Wind and Wave-Driven Circulation in a Fetch-Limited Coastal Basin: Western Long Island Sound

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A study of 8 years of observations from a buoy in western Long Island Sound shows that the wave field has an asymmetric response to wind direction. Waves are bigger when the wind blows from the east. Comparison to empirical models shows that the behavior is consistent with fetch limited wave growth when bottom dissipation is included.

Using 3 years of current observations at, and near, the surface, I show that the near surface shear also has an asymmetric response to wind. The shear is greater when the wind blows from the west. I then compare estimates of the near surface eddy viscosity for a range of wind stress values, of both signs, and show that the eddy viscosity is up to a factor of 5 greater when the wind is from the east.

A comparison of the variation of eddy viscosity coefficients with Henderson’s theory (Henderson et al., 2013), that describes the effect of waves on the eddy viscosity, shows that the observed response to wind directions is consistent with the predicted behavior. To assess the potential impact of surface intensification of the eddy viscosity on the wind driven flow in the estuary, I expand the model of Winant (2004) to include vertically variations in the eddy coefficient structure and use the finite element method to obtain solutions in the center of the domain. For geometries similar to that of Long Island Sound, I show that the vertical current structures are
much different from those which assume constant eddy viscosity coefficients and present evidence from current measurements that the variable eddy coefficient model provides a better representation of the flow.
Wind and Wave-Driven Circulation in a Fetch-Limited Coastal Basin:

Western Long Island Sound

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Chapter 1. Introduction

An understanding of near surface dynamics is of particular importance in the coastal ocean. Since surface water is the boundary between atmosphere and ocean, physical forces in the ocean boundary layer are key to explain how atmospheric factors influence the ocean over the vertical and seasonally. Also, the horizontal transport of human-induced water contamination affects marine environments and ecosystems. Therefore, quantitative assessments of surface water movement in response to wind and waves are environmentally and scientifically essential.

There are three major factors that interact with the near surface movement in estuaries; all posing substantial challenges to predicting surface water behavior. First, estuarine sediment transport is affected by the shape, size and variability of the watershed and associated river discharge (Wolman and Miller, 1960; Nash, 1994; Geyer et al., 2004). A clear power law relationship between sediment flux and river discharge has been established (Nash, 1994), which demonstrates the importance of the river discharge. Also, many biogeochemical processes, distribution of turbidity, dissolution of inorganic nutrients, particulate matters suspension the removal of nutrients by phytoplankton, are determined by mixing and stratification in an estuary (Harding et al., 1986; Cloern, 1987; Fisher et al., 1988; Kemp and Walter, 2012) and these are controlled by the pattern of surface salinity.

The second major factor affected by surface water dynamics is the transport of contaminants and nutrients from land. Nutrient enrichment induced by human activities can fuel algae growth and lead to hypoxia and harmful algae blooms (HABs). These are harmful to marine ecosystems and dangerous to humans (Anderson et al., 2002). On geological time scales, we see not only the occurrence of seasonal hypoxia in the coastal ocean or estuaries but also the area of
seasonal hypoxia tends to be increasing gradually as the eutrophication increases in many estuaries (Diaz, 2001). The occurrence and consequences of coastal eutrophication are strongly influenced by wind driven mixing and vertical stratification in the whole water column in shallow coastal water. (O’Donnell et al., 2004; Wilson et al., 2008; Scully, 2010a, b; Scully, 2013; O’Donnell et al., 2014; Wilson et al., 2015).

Finally and most importantly, the circulation in the whole estuarine water column is highly variable due to changing wind and wave conditions (Ekman, 1905; Wang, 1979; Gordon and Malcolm et al., 1987; Hunter and Hearn, 1987; Signell et al., 1990; Davies and Lawrence, 1995; Geyer, 1997; Winant, 2004; Whitney, 2011). The energy and momentum transport from the surface flow significantly affect both the horizontal and vertical motions.

There have been several observational and modeling studies of wind and wave driven circulation in the coastal ocean. Winant (2004) developed a barotropic wind-driven circulation model to study the effect of the shapes of the coastal basin. Polton et al. (2005), the analytical solutions of Eulerian velocity of Ekman component, Stokes component and Ekman-Stokes component with Hasselmann momentum equation. Lentz and Fewings et al. (2008) described how the onshore wave-driven transport near the surface compensates for the offshore undertow flow near bottom. Their observations and model results show that the offshore flow profiles result from the balance between the Coriolis force and the Hasselmann wave stress (created by the Stokes drift velocity influenced by the earth’s rotation). Terray et al. (1996) and Gerbi et al. (2009) showed the surface energy dissipation was enhanced by wave breaking in the open ocean and Henderson et al. (2013) developed a parameterization for the eddy viscosity in the surface layer which includes the effect of wave breaking energy dissipation. O’Donnell et al. (2014) discussed the observations of the winds and waves in the Long Island Sound using data from the LISICOS (Long Island Sound
Integrated Coastal Observation System) buoys, located at Execution Rocks (EXRK), Western Long Island (WLIS), Central Long Island Sound (CLIS) and Eastern Long Island Sound (ELIS). O’Donnell et al. (2014) showed that the characteristics of winds and waves in western Long Island Sound are quite different from the other parts of Long Island Sound and adjacent regions.

The goal of this study is to understand the interaction of winds, waves and surface currents in western Long Island Sound; a challenged and complicated estuary environment. I will make use of two observation systems: the CODAR (Coastal Ocean Dynamics Application Radar) and the ADCP (Acoustic Doppler Current Profiler). I will also use a model to assess the impact of the surface physics on the broader scale estuary flow. The modeling study uses a wind and wave driven barotropic semi-analytical model which includes depth-dependent eddy viscosity structured by the enhanced surface energy through wave breaking. This model is a modified form of Winant’s wind-driven analytical model (Craig and Banner, 1994; Terray et al., 1996; Burchard, 2001; Winant, 2004; Gerbi et al., 2009; Henderson et al., 2013).

I will test three hypotheses: 1) the combination of wind-fetch in coastal areas leads to larger wave amplitudes (significant wave height) when the wind is from the east in western Long Island Sound; 2) Near-surface shear in western Long Island Sound will be smaller when the wind is from the east due to wave effects on the near-surface eddy viscosity; and the wave enhanced eddy viscosity will 3) result in an asymmetry in near-surface current in response to wind.

The surface current data were collected from two CODARs (Coastal Ocean Dynamics Application Radar) which were at Great Captain Island, Greenwich CT and in Stehli Beach, Bayville, NY. Subsurface current data were observed from three bottom-mounted acoustic Doppler current profilers (ADCPs) deployed in western Long Island Sound. Wave data were
collected using a wave gage at the WLIS buoy as were wind speed and direction, barometric pressure and air temperature. The anemometer height was 3.5 meters above the water surface.

The chapters that follow describe the observations of wind, waves, surface and subsurface currents (Chapters 2, 3 and 4). Then, in Chapter 5, a wind and wave driven semi-analytical barotropic model will be used to help interpret these observations and to access implications of the observation. Chapter 2 will present annual and seasonal wind, wave and wind-fetch estimates in western Long Island Sound. The seasonal and annual observation results of surface currents, shears and eddy viscosity estimation are shown in Chapter 3. The thickness of Wave Affected Surface Layer (WASL), Wave breaking Layer (WBL), the asymmetric eddy viscosity and dissipation rate taking into account surface intensified turbulent energy dissipation are estimated using wind, wave and back scatter intensity of ADCPs in Chapter 4. Wind and wave driven semi-analytical model taking into account the surface intensified turbulent energy dissipation will be compared with the observations in Chapter 5.
Chapter 2. Waves in western Long Island Sound

2.1. Introduction

Long Island Sound (LIS) is a large estuary in southern New England that is separated from the northwest Atlantic by Long Island, NY. It is approximately 180 km in length and has a maximum width of 34 km. The bathymetry of LIS is shown in Figure 1. Most of the basin is shallower than 20 m, however, east of 72.5o W the sediments have been scoured and the depth is more typically 40m. A deeper east-west oriented channel also extends throughout the central part of the basin. The deepest areas of LIS are in the east where there is vigorous exchange of water with Block Island Sound through The Race. The western end of the basin, shown at higher resolution in the inset in Figure 1, is shallower (approximately 20 m and maximum depth is 70 m) and narrows towards the entrance to the East River. The water in WLIS is fresher as a consequence of exchange with the Hudson estuary and the discharge of freshwater from the many waste water treatment plants of the New York metropolitan area (O'Donnell et al. 2014). Deignan-Schmidt and Whitney (2018) show that Connecticut and Housatonic river water freshen western LIS in their modeling study. They show that the southern part of WLIS water is almost (~76%) from the Housatonic river water. Several observation and model studies (Bennet et al., 2010, O'Donnell et al. 2014, Deignan-Schmidt and Whitney, 2018) show the depth-average density in western LIS is approximately 25 psu and the water is fresher in Northern and Southern part of western Long Island Sound because of the location of rivers. Currents throughout LIS are largely driven by tidal forcing (Bogden and O'Donnell, 1997; Bennett et al., 2010) but both wind and waves intermittently make major modifications to the flow.
The bottom half of the water column in WLIS is hypoxic in the summer (O'Donnell et al., 2014) as a consequence of the eutrophication. Wilson et al. (2008) and O'Donnell et al. (2008a, b) have shown through data analysis and simulations that the near bottom oxygen concentration is greatly influenced by the surface wind stress through its modulation of both vertical mixing rates and the rate of restratification by straining of the longitudinal density gradient.

Three hundred years of coastal development has led to dense settlement along the shoreline of LIS and much of it is vulnerable to damage by waves during severe storms. The need to better understand the variability in coastal water quality and to improve design of coastal protection measures both motivate a better understanding of the variability in wind and wave conditions in LIS.

In western LIS (WLIS) the statistics of winds and waves have a directional asymmetry that is a consequence of the coastal geometry and regional meteorology (O'Donnell et al., 2014). Here we characterize the wind and wave statistics in more detail and examine whether the asymmetry can be explained by the fetch-limited wave parameterizations that have been widely used for several decades.

In the following section, we describe the data sources and the seasonal variation of wind and wave statistics. I then describe the relationships between wave heights and wind stress, and the direction of the wind.

2.2. Observation and method

Three ADCPs (acoustic Doppler current profilers) were located in the middle of across-Sound sections at the sites labeled EXRK, FB02 and WLIS in Figure 1. A buoy (WLIS buoy) was maintained at the WLIS site (40° 57.35’ N, 73° 34.8’ W) with wind speed and direction,
barometric pressure, relative humidity, and air temperature sensors (manufactured by R.M. Young Co.) at 3.5 m above the mean surface level. In addition, a Neptune Sciences, Inc. single axis non-directional wave module was operated. The WLIS buoy was 20 km from Execution Rocks, the eastern end of the East River, and 130 km from The Race, the eastern entrance to Long Island Sound. At the location of the WLIS buoy the Sound is approximately 10 km wide. Observations took place from 2007 to 2012. The spring season was defined as April to June; summer from July to September; autumn from October to December; and winter from January to March. Observations included wind magnitude, direction, peak wave period and significant wave height (the average height of the highest one-third of the waves during a 30 minutes sampling period). The wind data is converted to wind stress using COARE 3.0 (Fairall et al., 2003).

To calculate wind stress, bulk formulas are often used (Smith, 1980; Large & Pond 1981; Fairall et al., 1996; Fairall et al., 2003; Edson et al., 2013; Fisher et al., 2015) because the direct measurements of air-sea fluxes are challenging. These bulk parameterizations of wind stress are dependent on the drag coefficient and Monin–Obukhov (M-O) similarity theory (Monin & Obukhov, 1954; Obukhov, 1971). The Monin-Obukhov similarity theory assumes that there is a constant flux profile in the stationary and homogeneous ocean boundary layer.

The bulk parameterization of wind stress is

\[ \tau = -\rho_a C_D U_{10}^2 \]  

(1)

where \( \tau \) is wind stress, \( \rho_a \) is the air density, \( C_D \) is the drag coefficient and \( U_{10} \) is the wind speed at 10 m above the ocean surface. Since the wind speeds are measured at a height of 3.5m in WLIS, the wind speed is converted for the reference height of 10 m using the power law.
Figure 1. Bathymetry of Long Island Sound and research area in western Long Island Sound.
We assume that there is no stratification in the study area and so, the drag coefficient is

$$C_D = \left[ \frac{\kappa}{\ln \left( \frac{z}{z_0} \right)} \right]^2$$

(2)

The surface roughness length $z_0$ is partitioned into two parts; the viscous and rough turbulent components. The $z_{0, \text{smooth}} (= \gamma \frac{v}{u_*})$ represents the roughness length in the viscous smooth condition, and $z_{0, \text{rough}} (\alpha \frac{u^2}{g})$ accounts for the roughness due to the waves.

Then,

$$z_0 = \gamma \frac{v}{u_*} + \alpha \frac{u^2}{g}$$

(3)

where $\gamma$ is an empirical constant which is determined to be 0.11 (Edson et al., 2013), $v$ is the kinematic viscosity of air, $\alpha$ is empirical parameter known as the Charnock parameter. The Charnock parameter is a constant of 0.011 in the case of a fully developed sea in the open ocean (Smith, 1980, 1988; Large & Pond, 1981). In COARE 3.0 (Fairall et al., 2003) and COARE 2.5 (Edson et al., 2013), $\alpha$ is dependent on the wind speed. In fetch-limited environments, $\alpha$ is measured to be 0.0145 (Garratt, 1977), 0.018 (Wu, 1980), and 0.0288 (Geernaert et al., 1986). COARE 3.0 (Fairall et al., 2003) suggests $\alpha$ is dependent on significant wave height, wave period and wave speed (Taylor & Yelland, 2001 and Oost et al., 2002). To estimate wind stress, I used the various Charnock parameters to calculate the drag coefficient with Equation (2) for the neutral condition which can be utilized with the wind speed in Equation (1).

A comparison of the estimated wind stresses with the several different Charnock parameter $\alpha$ with the measured significant wave heights and wave period is shown in Figure 4 (Garratt, 1977;
Smith, 1980; Wu, 1980; Geernaert et al., 1986; Large & Pond, 1981; Fairall et al., 2003; Edson et al., 2013).

2.3. Wind in western Long Island Sound

The wind rose in Figure 2 shows the frequency of the wind stress magnitude (Pa) in direction bins using all data from 2007 to 2012. The different colors in the wind rose represent different wind stress ranges as defined on the right of the figure. It clearly shows that winds

Figure 2. Frequency distribution of the wind observations (wind rose) from the WLIS buoy using data from 2007 and 2012.
are more frequently from the northwest and southwest than other directions. However, there are also periods of large stress from the east.

There is substantial seasonal variation in the wind. Figure 3 (a) shows that during most of the winter the wind is from northwest and is from the southeast less than 10% of the time. The northwesterly stress magnitude is high and approximately 40% of northwesterly observations are larger than 0.05 Pa. In contrast, 80% of southeasterly wind stress observations are less than 0.05 Pa. Figure 3(d) shows that 44% of autumn wind observations are from the northwest and 40% of stress values are larger than 0.05 Pa. Southeasterly wind are weaker and 70% are less than 0.05 Pa. Winter and autumn wind patterns are similar with northwesterly winds are prevailing. High magnitude northwesterly winds predominate in both winter and autumn with a shift to the north in the winter. Figure 3 (b) and (c) show the wind is most frequently from southwest and northeast in the spring while the prevailing wind is largely from southwest in the summer. In the spring, the higher stress values occur when the wind is from northeast.

There are few studies of waves in Long Island Sound (Bokuniewicz and Gordon, 1980a and b; Signell et al., 2000; Rivera Lemus, 2008; O’Donnell et al., 2014). The along and across-Sound components of the wind are highly correlated and this complicates the relationship of winds and waves. To overcome these difficulties, I sorted the wave data in time to isolate periods when the across-Sound wind component magnitude is small, and then examine the response of the wave statistics to the along-Sound stress. I then performed an analogous procedure to examine the response to the across-Sound component. The thresholds were selected so that 75% of the observations exceed the value. The definition of critical values is based on the highest third of the wind which is significantly considered as high winds (Table 1).
Figure 3. Seasonal variation of the wind stress frequency distribution. (a) shows the winter (Jan-Mar), (b) spring (Apr-Jun), (c) summer (Jul-Sep) and (d) winter (Oct-Dec).

Table 1. Critical wind stress component values selected to identify periods when stress was directed along and across the axis of the Sound. 75% of observations exceed these thresholds.
2.4. Waves in western Long Island Sound

O'Donnell et al. (2014) show that the difference in significant wave height between western Long Island Sound and central Long Island Sound is larger when the wind is from the west than when it is from the east. This is a consequence of wave height in the western Sound being smaller than central Sound when the wind is from the west. I investigate whether this is a consequence of fetch.

In Figure 4, the eight different wind stress estimations are shown in the function of significant wave height ($H_s$) using the conditional bin average, i.e. I created the bins to divide the several ranges of data and each bin is averaged. The significant wave height is asymmetric in response to positive and negative along-Sound wind. The green and blue lines represent the classical wind stress estimations (Smith, 1980; Large & Pond, 1981; Smith, 1988), the magenta line represents wind speed dependent COARE 3.0 wind stress (Fairall et al., 2003), the black line is wind speed dependent COARE 3.5 wind stress (Edson et al., 2013), the black dashed line is wave dependent COARE 3.0 (Fairall et al., 2003), the green, blue and magenta dashed lines represent wind stress estimations in the fetch-limited environments (Garratt, 1977; Wu, 1980; Geenaert et al., 1986) and the red line with the error-bars represents the average wind stress of eight different formulations.

The horizontal error bars are the standard error computed with a decorrelation time of 20 hours. The significant wave height is most underestimated than the averaged wind stress using wave dependent COARE 3.0 method and the degree of asymmetry is greatest when the wind stress is calculated with wind dependent formulae in COARE 3.5. I found that the uncertainty of average wind stress is larger in negative along-Sound than in positive along-Sound wind especially when the wind magnitude is high.
Figure 4. Various wind estimations in the function of significant wave height (Hs).

Since the wave development is largely dependent on the wind characteristics, it is essential to understand wind characteristics to explain or to predict waves. In western LIS, wind correlation in the orthogonal component of wind is found as shown the blue line of which correlation coefficient is about 0.94 in Figure 5(a) and this correlation can make the wave and surface dynamics more complicated. Therefore, it is removed using the critical values in Table 1.

Figure 5(a) shows the correlation between along-Sound wind and across-Sound wind. The blue indicates the averages of the raw data, and the red represents the averages after periods when the across-Sound component exceeds the thresholds in Table 1 have been eliminated. After the data sorts out, the correlation coefficient is significantly decreased from 0.94 to 0.07.
Figure 5. (a) shows the across-Sound wind stress components averaged in bins of along-Sound stress. Note the disparity in scales. The blue line joins averages of the raw data, and the red line shows averages after periods when the across-Sound component exceeds the thresholds in Table 1 have been eliminated. The dashed blue line indicates the linear regression of the raw data before sorting out the data. (b) shows the significant wave height observed at WLIS, $H_s$, averaged in intervals of wind stress component magnitude. The blue and red lines show the dependence of $H_s$ on the along-Sound component before, and after, the larger across-Sound periods have been eliminated. The black and magenta lines show the dependence on across-Sound before and after the data sorting.
Figure 6. The dependence of $H_s$ on the components of the wind stress for (a) Jan-March, (b) April-June, (c) July-September, and (d) October-December. The color codes are as in Figure 5.

This decrease of correlation coefficient indicates that the wind correlation effect of orthogonal component on the wave is removed in order to examine the wave development in response to along-Sound and across-Sound wind separately. Figure 5(b) shows the dependence of $H_s$, averaged in bins of $\tau_x$, along-Sound wind stress, by the blue and red lines. The red line has the data from periods of high $\tau_y$, across-Sound wind stress eliminated. The difference between the blue and red lines is small suggesting that the effect of $\tau_y$ is not very significant when compared to $\tau_x$. $H_s$ is significantly smaller when $\tau_x > 0$ than when $\tau_x < 0$. The dependence of $H_s$ on $\tau_y$ is shown by the black and magenta dashed lines in Figure 5(b) with the magenta line showing
the averages after the periods of high $\tau_x$ have been eliminated. The difference between the black and magenta lines is substantial, particularly for $\tau_y > 0$, as a consequence of the greater sensitivity of $H_s$ to $\tau_x$. Note that the effect of the data sorting is to highlight that the response of $H_s$ to $\tau_y$ is symmetric.

The wave field in the western Sound is strongly correlated with the wind stress and that the response to winds from the west is approximately 50% of that due to the same stress from the east (Figure 6). In contrast, the response to wind stress from the north and south is symmetric. These observations are generally consistent with the dependence of the fetch at the WLIS buoy on direction. Though there is a strong seasonal variation in the statistics of the wind shown in Figure 3, Figure 6 shows that the dependence of the $H_s$ on $\tau_y$ is the same in all seasons.

2.5. Empirical Wave Models

The most fundamental theories for the generation of wave are described by Sverdrup and Munk (1947), Phillips (1957) and Miles (1957). Hasselmann et al. (1973) showed that the wave spectrum in a developing sea evolved through the application of mechanical work by the wind, transfer of energy to long wavelengths through nonlinear wave-wave interactions, and dissipation of energy by friction and breaking until, in a steady wind, a saturation spectrum is reached. This is known as a ‘fully developed sea’. Pierson and Moskowitz (1964) showed the wave spectrum in a fully developed sea for a range of different wind speeds. A ‘fully developed sea’ requires a very large fetch and long wind duration. If the wind duration and fetch are not enough, the wave field is said to be “partially developed”.

Numerical models to simulate the evolution of waves in more complicated forcing situations and realistic geometries have been developed. However, the evaluation of empirical
methods is warranted before more complex approaches are developed. The first empirical method was developed by Sverdrup and Munk (1947). Bretschneider (1957) and Wilson (1965) improved it with data collected in the 1950s. The Shore Protection Manual (U.S. Army, 1977), introduced the SMB (Sverdrup, Munk & Bretschneider) method for applications in coastal construction and marine transportation (Chue, 1977; Goda, 2003). In Lemus-Rivera (2008), he also revisited SMB equation to predict the significant wave height to compare with the observed significant wave height in central LIS. The SMB equation was modified with the empirical parameter to account for the local fetch.

Wilson (1965) showed that for the wind speed $u$, fetch length $F$, and gravitational acceleration $g$, the significant wave height, $H_w$, and period, $T_w$, can be described by the empirical formulae

$$H_w = 0.3 \frac{u^2}{g} \left\{1 - \left[1 + 0.004(gF/u^2)^{1/2}\right]^{-2}\right\}$$  \hspace{1cm} (4)

and

$$T_w = 1.37 \frac{2\pi u^2}{g} \left\{1 - \left[1 + 0.008(gF/u^2)^{1/3}\right]^{-5}\right\}$$ \hspace{1cm} (5)

For these parameterizations, the wind should be steady for a minimum time, $t_{min}$ (hours), where

$$t_{min} = F_{min}^{0.73} u^{-0.46}$$ \hspace{1cm} (6)

and minimum fetch should be at least $F_{min}$ (km), where

$$F_{min} = t_{min}^{1.37} u^{0.63}$$ \hspace{1cm} (7)

If $F > F_{min}$, then the wave growth is limited by the wind duration and $F_{min}$ should be used instead of $F$ in Equations (4) and (5) (Goda, 2003). Since the wind duration is seldom steady for more
than 20 hours Long Island Sound, F is smaller than $F_{\text{min}}$ ($F < F_{\text{min}}$) so that the wave growth is limited by the fetch.

Chue (1977) revisited the Bretschneider's summary of wave observations to develop an empirical formula that includes finite depth effects. Using the bottom friction factor $f$, water depth $d$, and the bottom slope $m$, Chue (1977) proposed

$$H_B = \frac{u^2}{g} H_w \tanh \left\{ \frac{1}{H_w u^2} \left( \frac{g}{0.16 + \frac{9.6}{(f/m)^{1.7}}} \left( \frac{gd}{u^2} \right)^{0.75 + \frac{1.64}{(f/m)}} \right) \right\}$$

(8)

and

$$T_B = \frac{u}{g} T_w \tanh \left\{ 6.66 \left( \frac{gd}{u^2} \right)^{0.41} \right\}$$

(9)

where the bottom friction factor was specified as, $f = 0.01$, based on observation in the Gulf of Mexico (Chue, 1977), the depth was assumed to be $d = 20 \text{ m}$ and the slope $m$ is about $10^{-4}$ at in the end of Long Island. $H_w$ and $T_w$ are computed from Equations (4) and (5). Chue (1977) also included the effect of the width, $W$, of the wave generating region in the parameterization using the observation results of Saville (1954) to compute an effective fetch, $F_{\text{eff}}$, as

$$F_{\text{eff}} = 1 - \exp \left( -\frac{1.24}{\sqrt{1 - \cos \theta}} \frac{W}{F} \right)$$

(10)

where $\theta = 30^\circ$ which is the angle of wind direction over which wind can be considered effective. The wind is limited to be effective on the coastal ocean because of complicated geography known as a wall effect.
As it is clear in Figure 1, the fetch at the WLIS site in western Long Island Sound, is much longer when the wind is from East. We can, therefore, test whether the fetch limitation as expressed in Equations (4) to (9) can explain the asymmetry in the wave height to winds stress shown in Figures 4 and 5 as was speculated in O'Donnell et al. (2014). Figure 8 (a) shows the significant wave height, $H_s$ (m) measured at WLIS, sorted to eliminate intervals with large $\tau_y$ and averaged in intervals of $\tau_x$ by the black symbols and lines. $H_s$ increases with the magnitude of the stress irrespective of the sign, however, for negative values (wind from the East) the rate is approximately twice that of positive values. At the largest magnitude of the wind stress components where data were available, $\pm 0.3 \ Pa$, easterly winds lead to $H_s$ in excess of 1.5m, but for westerly winds $H_s$ only reaches 0.8m. This appears to be consistent with the expectation from the empirical models since the fetch from the east is an order of magnitude larger than that from the west. The solid blue lines represent the Wilson parameterization, $H_B$, as expressed in Equation (4) to (7) and computed with $F = 56 \ km$ for $\tau_x < 0$, and $F = 14 \ km$ for $\tau_x > 0$. These values of fetch appear to be in good agreement with the observations. The solid red lines show the predictions of the Chue (1977) revision of the Bretschneider's parameterization, $H_B$, as expressed in Equations (8) and (9). The

<table>
<thead>
<tr>
<th></th>
<th>W method (Best fit)</th>
<th>W method with GF</th>
<th>B method (Best fit)</th>
<th>B method with GF</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fetch(km) in $\tau_x$</td>
<td>$\tau^-$ 56 $\tau^+$ 14</td>
<td>$\tau^-$ 120 $\tau^+$ 10</td>
<td>$\tau^-$ 200 $\tau^+$ 14.5</td>
<td>$\tau^-$ 120 $\tau^+$ 10</td>
</tr>
<tr>
<td>Fetch(km) in $\tau_y$</td>
<td>$\tau^-$ 13.5 $\tau^+$ 13</td>
<td>$\tau^-$ 10 $\tau^+$ 10</td>
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Table 2. The estimated fetch lengths in each with Wilson’s parameterization (W) for best fit, the Wilson’s parametrization with Geographic Fetch (W with GF), the Bretschneider’s parameterization (B) for best fit and Bretschneider's parameterization with Geographic Fetch (B with GF) in along-Sound wind ($\tau_x$) and in across-Sound wind ($\tau_y$). The $\tau^-$ represents the negative wind stress and $\tau^+$ indicates the positive wind stress for each wind event.

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local geographic fetch lengths are $F = 200 \ km$ for $\tau_x < 0$, and $F = 14.5 \ km$ for $\tau_x > 0$. Again, these provided a reasonable representation of the variation in the data.

The local geographical fetch is estimated based on the geometry of the basin and the location at the WLIS buoy location. The distance from the race of Eastern LIS to WLIS station is about 120 km and the distance from East river to WLIS station is about 10 km. Therefore, the geographic fetch is 120 km when the wind is from the east and 10 km when the wind is from the west. The blue dashed line in Figure 8 represents Wilson’s method using the geographic fetch. The red dashed line represents Bretschneider’s method. The solid lines show the parametrizations with fetch closer to provide a best-fit to the observations. The best-fit values in the Bretschneider’s formula are consistent with those based on geography. However, the Wilson’s formula is inconsistent. The superior performance of the Bretschneider’s method implies that dissipative processes are significant.
Figure 8. (a) The significant wave height, $H_s$, measured at WLIS, sorted to eliminate intervals with large $\tau_y$, and averaged in intervals of $\tau_x$ are shown by the black symbols and lines. The empirical relationships of Wilson and Bretschneider are shown by the dashed and solid lines respectively. (b) shows the dependence of $H_s$ on the $\tau_y$, together with the empirical predictions. W method indicates Wilson's parameterization (blue) and B method indicates Bretschneider's parameterization (Red). The dashed lines represent the predicted $H_s$ with local distance (Geographic Fetch) from the land. As each of the associated fetch values is summarized in Table 2.

Figure 8(b) summarizes the variation of the observed $H_s$, averaged in intervals on the across-Sound wind stress, $\tau_y$, after periods of high along-Sound stress have been eliminated. Note
that the $H_s$ scale is different from those in Figure 8(a). The dependence of $H_s$ on $\tau_y$, is almost symmetric. At the highest stress values $H_s$ is approximately 1m, which is consistent with the positive side of the $\tau_x$ dependence shown in Figure 8(a). The empirical predictions of the two used methods are almost the same using optimal choices of fetch. $F = 13 \text{ km}$ for $\tau_y > 0$ and $F = 13.5 \text{ km}$ for $\tau_y < 0$. Both models are consistent with the observations with suitable choices for the fetch. These are compared with the red dashed lines with the local fetch length which is 10 km both in positive and negative across-Sound wind.

To find the best fit to observation, the minimum values of RMS (Root Mean Square) and maximum values of slope of Robust Fit regression are used. When using fetch values estimate from the geography, the empirical prediction with Wilson’s method clearly fails to represent the trend in the observation. Bretschneider’s method, however, a good agreement with observations.

2.6. Summary and Conclusion

I have presented a summary of the annual and seasonal variability of the wind speed and direction in western Long Island Sound. Figure 2 shows that the wind is most frequently from the west and during periods of higher stress (0.1 Pa), it is most frequently from the north. Wind statistics show clear seasonal differences. Figure 3 shows that in the fall (October to December) and winter (January to March), the wind is mostly from northwest and west and is frequently in the range 0.09-0.1 Pa. Winds are from the southwest and east in the spring (April to June) and are most frequently from the southwest in the summer (July to September). Comparison to previous studies (Lentz et al. 2008; O'Donnell et al. 2014) shows that the magnitude of the wind stress in western Long Island Sound is less than in the central Sound. The wind direction statistics are similar, though in the winter the western Sound experiences more frequent winds from the north and more from the west in the summer.
I have also presented a summary of the annual and seasonal dependence of the significant wave height in the western Sound on wind stress. After removing the correlation between along-Sound and across-Sound wind stress components by sorting the data to eliminate periods when the wind was not aligned along, or across, the Sound, I find that there is an asymmetric variation of significant wave heights with the along-sound wind stress and a symmetric variation with the across-Sound components. This pattern is a consequence of the asymmetric wind-fetch in the along-Sound regime, i.e. when the along-Sound wind is negative, the wind-fetch length is greater than when the along-Sound wind is positive. These results are consistent with those in O’Donnell et al. (2014) who compared waves in the western and central parts of the Sound.

Finally, I quantitatively assess the performance of well-established wave height parameterizations on the fetch to the observations. I find that Chue (1977) adapting at the Bretchneider (1957) model is consistent with the observations in that they predict the symmetric response to the across-Sound wind stress component and the asymmetric response to the along-Sound stress component. This is true for all seasons. However, with fetch values for winds from the east chosen from the geometry of the basin, the significant wave height is significantly over-predicted in Wilson (1963)’s parameterization while it is well predicted in Bretschneider’s parameterization which takes into account the bottom dissipation and shallow water effect.

In conclusion, the wave field in western Long Island Sound is primarily influenced by the local wind and can be predicted by empirical models that assume the waves are fetch limited. However, it is clear that dissipative processes are significant and must be represented.
Chapter 3. The role of wind and wave in the near surface dynamics in Western Long Island Sound

3.1. Introduction

Understanding the dynamics of near surface currents is essential to predicting the water movement and transport of materials in the coastal ocean and estuaries (Wolman and Miller, 1960; Nash, 1994; Geyer et al., 2004). Skillful simulations are required to predict coastal sediment transport (Nash, 1994), biogeochemical processes (Harding et al., 1986; Cloern, 1987; Fisher et al., 1988; Kemp and Walter, 2012) and to understand contaminant distributions and the processes leading to eutrophication (Anderson et al., 2002; Diaz, 2001). In estuaries, tides, mixing and stratification (O’Donnell et al., 2004; Wilson et al., 2008; Scully, 2010a, b; Wilson et al., 2012; Scully, 2013; O’Donnell et al., 2014) and wind (Ekman, 1905; Wang, 1979; Gordon and Malcolm, et al., 1987; Hunter and Hearn, 1987; Signell et al., 1990; Davies and Lawrence, 1995; Geyer, 1997; Winant, 2004; Whitney and Codiga, 2011) have all been shown to play important roles in determining the circulation.

Most studies of estuarine circulation have examined how the interaction of the tidally imposed pressure gradients influences the longitudinal (along-channel) and lateral (across-channel) circulation and lead to, and are modified by, vertical mixing (Simpson et al., 1990, 1991; Sharples et al., 1994), and wind (Weisberg, 1976; Elliott, 1978; Wang, 1979; Noble et al., 1996; Geyer, 1997; North et al., 2004; Scully et al., 2005; Chen and Sanford, 2009). The important role of the Earth’s rotation, Ekman dynamics, has also received considerable attention in larger estuaries (Malone et al., 1986; Sanford et al., 1990; Geyer et al., 2001; Lacy et al., 2003; Winant, 2004; Lerczak and Geyer, 2004; Wilson et al., 2008; Scully et al., 2009; Scully, 2010; Li and Li, 2011).
The effects of wind in estuaries has received less attention and, until recently, studies have focused on the interactions between the wind stress and the barotropic pressure gradient (Wang, 1979; Geyer, 1997; Garvine, 1985; Scully, 2005; Chen and Sanford, 2009; Whitney and Codiga, 2011). However, the wind also affects the ocean surface water through the generation of high frequency surface gravity waves, with periods less than 8s in Long Island Sound (O’Donnell et al., 2014). Though there has been considerable recent work on how momentum is transported through waves to turbulence in the surface boundary layer (Janssen, 1989; Craig and Banner, 1994; Terray et al., 1996; Gerber et al., 2009), only recently have studies addressed the role of the higher frequency components of the wind stress on the modification of the vertical shear and vertical mixing in estuaries (O’Donnell et al., 2008; Wilson et al., 2008; Scully et al., 2009; Scully, 2010; Li and Li, 2011, 2012; Henderson et al., 2013).

This chapter describes an analysis of observations of the near surface dynamics in western LIS where the interaction of advection and mixing is critical to understanding the variability of hypoxia (O’Donnell et al., 2008). I demonstrate that the vertical shear in the horizontal currents near the surface is much larger when the wind stress is from the west than from the east. Using the observed stress and shear, I then investigate the effective near surface eddy viscosity computed using the gradient-flux relationship.

Section 3.2 describes the important geographic characteristics of western Long Island Sound and the location of the observations. It also describes the character of the Coastal Ocean Dynamics Application Radar (CODAR) and Acoustic Doppler Current Profiler (ADCP) current measuring systems, and then aspects of the data processing. In Section 3.3, the surface current data procession is described with surface Stokes drift. In Section 3.4, the near surface shear is calculated and shown to have an asymmetric response to the direction of the wind. In Section 3.5, the near
surface eddy viscosity is estimated using measurements of the shear and stress. The results are summarized and discussed in Section 3.6.

3.2. Observation setting

The geography of western LIS is shown in Figure 1 of Chapter 2. To measure the circulation, I used two different instrument systems, high frequency RADAR (CODAR) and bottom mounted ADCPs. The CODAR sites were located at Great Captain Island (GCAP), Greenwich, CT, and at Stehli Beach (STLI), Bayville, NY. The ADCPs were located at the middle of LIS at the sites labeled EXRK, FB02 and WLIS in Figure 1 of Chapter 2. Waves were measured at WLIS and meteorological instruments were maintained buoys at WLIS and EXRK. I use data spanning the interval 2007 to 2009. Locations are summarized in Table 1.

CODAR systems illuminate the sea surface using high frequency RADAR and use the Doppler shift in the backscattered radiation from surface gravity waves backscattered via the Bragg mechanism to estimate the current magnitude (Crombie, 1973). The Doppler shift leads to an estimate of the ocean surface current in the direction along a line between the transmitter and the scattering waves. Two or more sites are, therefore, required to obtain the vector current (Paduan and Hans, 1997; Teague et al., 1997). The spatial resolution of CODAR at the STLI and GCAP is approximately 1 km, and the maximum range is 35 km. The current that is measured is effectively an average over the surface 50cm of the ocean (Stewart and Joy, 1974; Emery, 2004). Figure 1 illustrates the spatial coverage and resolution of two CODARs.

The ADCPs measure the sub-surface currents in most of the water column using the Doppler shift in the acoustic backscatter from Sound pulses propagating upwards from the instruments on the bottom. Measurements can’t be obtained near the instrument since the
transducers take time recover from transmitting the outgoing pulse (Mueller et al., 2007). This “blanking distance” depends on the ADCP frequency and was 1.76m from the ADCP to the midpoint of the first good data bin at WLIS and FB02. As a consequence of the acoustic transducer design and the beam geometry, velocity also can’t be estimated in a layer below the surface of thickness approximately equal to 10% of the distance from the ADCP to the surface (Mueller et al., 2007). In addition to velocity, ADCPs also record the acoustic backscatter intensity variation with range and the pressure at the instruments. These can be used to determine the level of the water surface.

Figure 1. The locations of the CODAR sites at (a) Great Captain Island, Greenwich CT, (G.C.) and (b) Stehli Beach, Bayville, NY, (STLI). The locations of the ADCPs are indicated by EXRK, FB02 and WLIS. The blue dots are the locations of the CODAR current estimates.

Together, these instruments provide estimates of the velocity over most of the water column at the locations of the ADCPs. I use the difference between the CODAR derived near surface current and the velocity in the top bin of the ADCP, divided by the separation of the centers
of the bins to estimate the surface shear. The top usable bin is located using the pressure sensor record and follows the sea surface level fluctuation. I also define the sub-surface shear as the difference between the velocity in the top ADCP bin and that two bins below.

<table>
<thead>
<tr>
<th>Deployment date</th>
<th>Measurements</th>
<th>Station</th>
<th>Latitude (°N)</th>
<th>Longitude (°W)</th>
<th>Frequency and Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>2007 - 2009</td>
<td>ADCP (Acoustic Doppler Current Profiler)</td>
<td>WLIS</td>
<td>40.9541</td>
<td>73.5790</td>
<td>300 kHz with 1m bins</td>
</tr>
<tr>
<td></td>
<td></td>
<td>FB02</td>
<td>40.9253</td>
<td>73.6565</td>
<td>300 kHz with 1m bins</td>
</tr>
<tr>
<td></td>
<td>CODAR (Coastal Ocean Dynamics Application Radar)</td>
<td>Great Captain</td>
<td>40.9820</td>
<td>73.6237</td>
<td>25.3MHz with 1km resolution</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Stehli Beach</td>
<td>40.9087</td>
<td>73.5873</td>
<td>26.19MHz with 1Km resolution</td>
</tr>
</tbody>
</table>

Table 1. Details of ADCP and CODAR deployment

There are several earlier studies of the root mean square (RMS) difference between ADCP and CODAR measurements (Holbrook and Frisch, 1981; Leise, 1984; Matthews et al., 1988; Shay et al.; 1995, Chapman et al., 1997; Graber et al. 1997). Graber et al. 1997 showed that the contribution of measurement uncertainty in CODAR and ADCP data to the RMS difference is approximately 20~30%, consequently, the near surface shear must account for the remaining 70-80%. Recently, Hubbard et al. 2013 estimated the uncertainty in CODAR measurements was less than 0.1~5% of the mean current.
In this study, I use the hourly vector velocity data from the CODAR cell nearest to each ADCP location and average the ADCP data at the top useful bin to 1 hour intervals. Both CODAR and ADCP vector series are rotated to an along-Sound (60°) and across-Sound (330°) coordinate system using

\[ u_{along} = u \cos \theta + v \sin \theta \]

\[ v_{across} = -u \sin \theta + v \cos \theta \]

where \( \theta = 30^\circ \). The surface shear is then estimated from the difference between the CODAR velocity estimate and the velocity in the top useful ADCP bin.

I use wind velocity observations and surface air and water temperature measurements from the WLIS buoy to compute the wind stress using the COARE 3.0 bulk formula (Fairall et al., 2003). The stress vector series are rotated to the along-Sound, \( \tau_x \) and across-Sound, \( \tau_y \), frame using Equations 1a and 1b.

3.3. Current data processing

In western LIS the presence of the East River, and the along-Sound density gradient it creates, causes the sub-tidal average along-Sound surface current to be directed eastward (Wilson 1976; Ianniello 1977a, b; O’Donnell and Bohlen 2003; Crowley 2005; Hao 2008; Bennet 2010; O’Donnell et al 2014). Although there have been several studies that show the importance of wind to estuarine circulation (Weisberg 1976; Elliott 1978; Wang 1979; Noble et al. 1996; Geyer 1997; Scully et al. 2005; Chen and Sanford 2009), the wind and wave effect on estuarine circulation in western LIS is less studied. Existing observations and simulations (O’Donnell et al. 2008; Wilson et al. 2008; O’Donnell et al. 2014; Wilson et al. 2015) show that the wind modulates the vertical density gradient, the circulation, and the variability of dissolved oxygen near the bottom.
It is well established (Stewart and Joy, 1974; Teague, 1986; Fernandez et al., 1996; Paduan and Rosenfeld, 1996; Graber et al., 1997; Laws, 2001; Ullman et al., 2006; Mao and Heron, 2008; Ardhuin et al., 2009; Kirincich et al., 2012) that the mean velocity created by surface gravity waves, or Stokes Drift, influences the velocity estimated by HF RADAR current measuring systems. To compute the Eulerian mean velocity and shear, the magnitude of the effect must be estimated and removed. Recently, Chavanne (2018) summarized the methods that have been developed to account for the wave effects and compared the magnitude of corrections necessary at a range of wind speeds assuming that the wave spectrum was saturated and described by the Philips (1958) spectrum. He also summarized predictions for the effective depth that the HF RADAR derived currents represent.

Stewart and Joy (1974) showed that the phase velocity of a surface gravity wave at the Bragg wavelength propagating on a steady, vertically sheared current with velocity $u_E(z)$ will be estimated by an HF RADAR as $c = c_0 + \Delta c_E$, where $c_0$ is the phase velocity from linear wave theory (Stokes, 1847) and

$$\Delta c_E = \int_{-\infty}^{0} u_E(z) \exp(2k_Bz) \, dz,$$

(2)

i.e. a weighed vertical average with weights that decay with depth in proportion to the magnitude of the Stokes (i.e. the Lagrangian mean particle velocity) drift due to the Bragg wave.

Stewart and Joy (1974) also conjectured that in the absence of a Eulerian mean flow, $u_E = 0$, the HF RADAR systems would measure the Stokes drift induced by the Bragg wave, $\Delta c_B$, with wavenumber $k_B$ and amplitude, $a_b$, averaged as in Equation (2), which can be written as
\[
\Delta c_B = \frac{c_0}{2} k_b^2 a_b^2
\]  

or one half of the Stokes drift velocity at the surface, \( u_B(0) \), created by the wave train. They noted that this is precisely the shift in phase speed predicted by Stokes for the weakly non-linear self-interaction of the finite amplitude Bragg wave. Laws (2001) followed this approach and showed that if the Stokes drift due to a more complex unidirectional wave field with spectrum \( S(k) \) was viewed as an Eulerian flow and included in Equation (2), then the adjustment to the phase speed of the Bragg wave would be

\[
\Delta c_L = 2k_B \int_0^\infty \omega(k) \left( \frac{k}{k + k_B} \right) S(k) dk
\]

where \( \omega(k) \) is the angular frequency. Chavanne (2018) pointed out that it is unclear why the wave train would be translated by the mean particle velocities created by the waves. However, Longuet-Higgins and Phillips (1962), and Huang and Tung (1976) argued that wave-wave interactions could directly modify the phase velocity of the Bragg waves. For a continuous, unidirectional spectrum, \( S(k) \), Longuet-Higgins and Phillips (1962) considering only resonant interactions and showed the phase speed shift would be

\[
\Delta c_{LH} = 2 \int_0^\infty \omega(k) k S(k) dk + 2k_B \int_0^\infty \omega(k) S(k) dk
\]

Later Huang and Tung (1976) included non-linear interactions across the spectrum and concluded
\[ \Delta c_{HT} = \int_0^\infty \omega(k)kS(k)dk \]  \hspace{1cm} (6)

which is again one half of the Stokes drift at the surface due to the entire wave spectrum. Note that though the Huang and Tung (1976) and Longuet-Higgins and Phillips (1962) results can be expressed in terms of the Stokes drift, the central argument is that the wave dynamics modify the phase speed of the Bragg waves, and not that the Stokes velocity is translating the wave field.

Chavanne (2018) illustrated the magnitude of the velocity adjustments in Equations (2-6) for a range of wind speeds for a 13.5 MHz HF RADAR system. He showed that the smallest expected shift in the phase speed of the Bragg wave train was that predicted by the Huang and Tung (1976) theory, i.e. one half of the surface Stokes velocity. Neglecting some wave interactions (Longuet-Higgins and Philips, 1962), or treating the Stokes velocity as an Eulerian mean flow (Stewart and Joy, 1974; Laws, 2001), increases the velocity shift by up to 175%. Chavanne (2018) concluded that though the results of Huang and Tung (1976) were physically appealing, additional experimental investigations were necessary to resolve the role of waves on HF RADAR estimates of surface currents.

Since wave field in LIS is fetch limited, and the CODAR systems in LIS are higher frequency (25 MHz), I estimate the effect of waves on the current using the JONSWAP (Hasselmann et al., 1973) wave spectrum (with peakiness factor \( \gamma = 1 \)) as

\[ S(\omega) = \frac{\alpha}{\omega^5} g^2 \exp \left[ -\frac{5}{4} \left( \frac{\omega_p}{\omega} \right)^4 \right] \]  \hspace{1cm} (7)

where \( \alpha = 0.076 \left( \frac{\mu^2}{Fg} \right)^{0.22} \) and \( \omega_p = 22 \left( \frac{g}{\mu_{1/0}F} \right)^{1/3} \), where the fetch values (\( F \)) are as estimated in
Figure 2 shows the predicted correction to the Bragg wave phase velocity at the WLIS site as a function of wind stress using equations (6) and (7) (the Huang and Tung (1976) approach), as the blue line. These values are equivalent to one half of the Stokes drift at the surface due to the wave field and form the lower bound of the range in the analysis of Chavanne (2018). The red line shows the predicted dependence of the surface Stokes drift on wind stress for comparison. The velocity estimates I acquired from the CODAR systems in western LIS are corrected by subtracting $\Delta c_{HT}$, i.e. the values shown in blue in Figure 2. To provide an upper bound on the potential error introduced I will also show the consequences of subtracting $2\Delta c_{HT}$.

The current vector, $\mathbf{u}$, estimated by the CODAR (corrected) and ADCPs and can be decomposed into the tidal part $\mathbf{u}_t$, the steady estuarine circulation, $\mathbf{u}_e$, and the residual variation, $\mathbf{u}_w$, which I attribute to wind and wave-driven motion. This is expressed as

$$\mathbf{u} = \mathbf{u}_t + \mathbf{u}_e + \mathbf{u}_w \quad (8)$$

The tidal component, $\mathbf{u}_t$, of the current is estimated using harmonic analysis (Pawlowicz et al., 2002) and I include the constituents $M_2, M_4, M_6, S_2, N_2, K_2, K_1, O_1$ and $P_1$. The estuarine flow, $\mathbf{u}_e$, is estimated as the record mean at each depth level. The residual, $\mathbf{u}_w$, is then rotated into the along, and across Sound components $U_w, V_w$. 
Figure 2. The blue line shows the dependence of $\Delta c_{HT}$, the Huang and Tung (1976) correction to the Bragg wave phase speed, on the along-Sound wind stress at the WLIS buoy. This is equivalent to one half of the surface Stokes drift. The red line shows $2\Delta c_{HT}$.

To simplify the interpretation of the observations I partition the velocity data records into three sets based on the sign of the wind stress components. The data set $U_{w}^{T_{x}+}$ contains all estimates of the along-Sound wind driven velocity component obtained when the along-Sound wind stress component was positive ($\tau_x > 0$) and the across-Sound component is small ($|\tau_y| < \tau_{yCR}$). Analogously, $U_{w}^{T_{x}-}$ are the estimates when the along-Sound stress was negative ($\tau_x < 0$) and...
$|\tau_y| < \tau_{yCR}$. The remaining data were acquired during periods when both the along and across-Sound components were significant and these data were ignored.

In Figure 3, the vertical structure of the mean along Sound wind driven current in the top 5m of the water column estimated by both CODAR and ADCPs when the along Sound wind stress is positive, $< u^x_+ >$, is shown in red. The mean structure during intervals of negative wind stress, $< u^x_- >$, is shown in blue. The horizontal lines show the magnitude of the 68% confidence interval (two standard errors) using a decorrelation time estimate of 20 hours. The uncorrected CODAR estimates are shown at a depth of -0.5m by the red and blue circles. The diamonds show the mean after subtracting $\Delta c_{HT}$ from the CODAR velocity series, and the squares show the consequence of a $2 \Delta c_{HT}$ correction. These corrections span the range of plausible values necessary to remove the effects of the surface gravity waves on the phase velocity of the Bragg waves that backscatter the CODAR. For both signs of the wind stress the magnitude of the mean near surface current is reduced, however, the magnitude of the correction is larger for $\tau_x < 0$. Clearly, the mean near surface wind driven flow is downwind at approximately 0.02 m/s. This is comparable to the long term mean flow (not shown) that we attribute to the estuarine circulation.
Figure 3. The vertical structure of $u_m^{T_x^+}$ (red dashed line), $u_m^{T_x^-}$ (red solid lines), $u_m^{T_y^-}$ (blue dashed line) and $u_m^{T_y^+}$ (blue solid lines). The depth levels are relative to the moving surface and are 0.5m, 2.75m, 3.5m, 4m, 4.5m and 5m.

3.4. Surface shear and wind

I use the difference between the CODAR and ADCP measurements divided by the distance between the level of the CODAR measurement (0.5 m) and that of the uppermost useable ADCP bin, 2.75 m below the surface, to estimate the near-surface vertical shear in the horizontal wind driven velocity components: i.e., $S_{w}^{T_x^+/−} = (U_{wCODAR}^{T_x^+/−} − U_{wADCP}^{T_x^+/−})/2.25$ and $S_{w}^{T_y^+/−} = (U_{wCODAR}^{T_y^+/−} − U_{wADCP}^{T_y^+/−})/2.25$. To estimates the wind stress, I use the COARE 3.0 formula (Fairall et al., 2003) and observations of the wind velocity components, air temperature and humidity obtained at the WLIS buoy. The black line in Figure 4 shows the vertical shear in the along-Sound
current component estimates, $S_{\tau_x^+}$ and $S_{\tau_x^-}$, based on the uncorrected CODAR velocity estimates, averaged in bins of the along-Sound wind stress component $\tau_x$. The horizontal and vertical error bars show the 68 percent confidence interval surrounding the bin-mean stress and shear respectively.

![Figure 4](image)

Figure 4. The variation of the bin average near surface shear in the along-Sound current with the along-Sound wind stress. The shading represents the distribution of the raw data.

In computing the confidence intervals, I again assume that the decorrelation time scale for the data series is 20 hours. The data distribution is illustrated by the grey shading which represents the percentage of all the samples that fall in each 0.025 Pa by 0.025 s$^{-1}$ bin. The scale is provided on the right of the figure. Most of the data show low wind stress and shear magnitudes, however, at higher stress magnitudes the magnitude of the shear is much larger when the wind stress component is positive, than when negative.
The black line in Figure 5 shows the same results as in Figure 4 with the uncertainty of about 65%. To demonstrate the impact of the corrections to the CODAR current estimates due to the wave field, the blue line shows the bin-averaged shear computed after $\Delta c_{HT}$ has been subtracted from the CODAR velocity estimates, and the red line shows the consequence of a $2 \Delta c_{HT}$ correction. Since as shown in Chapter 2, the fetch at the WLIS buoy is much larger when the wind is from the east ($\tau_x < 0$), the waves heights are also larger, consequently, the adjustment to the direct CODAR estimates are more significant, approximately 60% of the direct CODAR estimate. It is clear from Figures 4 and 5 that the sign of the wind stress plays an important role in the magnitude of the near surface shear. A stress of 0.15 $Pa$ from the west creates approximately twice the vertical shear than the same magnitude stress due to a wind from the east.
Figure 5. Bin averaged near surface shear dependence on the along-Sound wind stress. The black line shows the shear based on the uncorrected CODAR measurements and the blue and red lines the effect of the $\Delta c_{HT}$ and $2\Delta c_{HT}$ corrections. The dashed lines show linear regressions through the shear values for positive and negative wind stress components.

3.5. Estimation of Eddy viscosity

The conventional gradient-flux eddy viscosity representation of turbulent momentum transport in a boundary layer can be expressed as (Thorpe, 2005)

$$A_x = \tau_x / (\rho \partial u / \partial z)$$

where $\tau_x$ is along-Sound wind stress, $\partial u / \partial z$ is the vertical shear in the current, $A_x$ is the eddy viscosity and $\rho$ is the water density. It is well established that the eddy viscosity coefficient is determined by both the flow and the boundary geometry and many models have been proposed for
the estimation of $A_z$. Since few field measurements are available in situations where wind and gravity waves are important, I use the estimates of the bin-averaged wind stress and near surface shear shown in Figure 5, and Equation (10) to estimate the vertical eddy coefficient as $\frac{1}{\rho \beta}$ where $\beta$ is the slope of the dashed lines in Figure 5.

![Figure 6. The estimated surface eddy viscosity. It depicts the eddy viscosity in the along-Sound wind and represents $A_z$ after taking out $\Delta c_{HT}$ correction. The dashed lines represent the averaged values for positive and negative stress.](image)

For $\tau_x > 0$ the effect of the wave field correction to the CODAR velocity estimates are small and the data are consistent with an eddy viscosity of $A_z^+ = 2.5 \times 10^{-3} \ m^2/s$. In contrast, for the winds from the east ($\tau_x < 0$) the effect of the correction for waves is considerable and decreases the shear. The slope of the blue dashed line in Figure 5 (the $\Delta c_{HT}$ correction) implies an
eddy viscosity $A_x^* = 10^{-2} \text{ m}^2/\text{s}$ and the black and red dashed lines imply values of $5 - 25 \times 10^{-2} \text{ m}^2/\text{s}$.

An alternative approach to estimating the magnitude of the eddy viscosity would be to divide the bin averaged stress by the shear $(\frac{\tau_x}{\rho} / \frac{\partial u}{\partial z})$ and Figure 6 shows the results using the data shown in blue in Figure 5 (the $\Delta c_{HT}$ correction). The means for easterly ($7 \times 10^{-3} \text{ m}^2/\text{s}$) and westerly ($4 \times 10^{-3} \text{ m}^2/\text{s}$) wind components are shown by the dashed lines. Both approaches agree that the eddy viscosity is enhanced when the wind or from the east ($\tau_x < 0$) than from the west.

### 3.6. Summary and Conclusions

There are several important results in this paper. First, the vertical shear in the wind forced surface downwind currents are asymmetric in response to oppositely driven winds. This is because the surface currents are largely influenced by not only wind but also wave dynamics. As several recent studies show the importance of wave-induced turbulence on estuarine surface water (Scully et al., 2015; Scully et al., 2016; Fisher et al., 2017) and surface intensified turbulence through the wave breaking (Janssen, 1989; Craig and Banner, 1994; Terray et al., 1996; Gerber et al., 2009; Henderson et al., 2013), wind-driven surface currents should be modified by the wave in the coastal ocean. In a fetch-limited coastal basin such as western Long Island Sound, the wave dynamics are shown to be asymmetric in response to wind direction (Ch. 2). This results in the asymmetric surface turbulent energy which makes the surface currents and surface shears asymmetrically.

Second, the larger $\Delta c_{HT}$, which is equivalent to Stokes drift, occurs along with the long fetch in the negative along-Sound wind rather than the shorter fetch in the positive along-Sound wind. This $\Delta c_{HT}$ is measured in CODAR surface current while ADCP estimation doesn’t include (Teague, 1986; Fernandez et al., 1996; Paduan and Rosenfeld, 1996; Graber et al., 1997; Laws,
Therefore, the surface $\Delta c_{HT}$ (Chavanne, 2018) is taken out from the CODAR surface estimation to make the surface shears in response to wind. The results show that the $\Delta c_{HT}$ correction makes the surface shear reduce about 50% in the negative along-Sound wind and 10% in the positive along-Sound wind.

Third, as a consequence of the asymmetric surface shears, the mixing rate quantified as the eddy viscosity is also asymmetric. When the wind is negative along-Sound, the surface shear is small and the eddy viscosity is large while the surface shear is large and the eddy viscosity is small when the wind is positive along-Sound. This indicates that the surface water is more intensely mixed over the vertical by negative along-Sound winds than by positive along-Sound winds. This asymmetry of surface eddy viscosity with the asymmetry of surface shears also results from the asymmetric wave dynamics in western LIS. When the wave is larger in negative along-Sound, the surface shear is smaller and the surface eddy viscosity is larger because of more surface turbulent energy while the surface shear is larger and the surface eddy viscosity is smaller in the positive along-Sound wind.

I also found that the surface $\Delta c_{HT}$ correction makes the surface eddy viscosity increase largely. The eddy viscosity increases about twice in the negative along-Sound wind and increases about 1.5 times in the positive along-Sound wind after controlling the $\Delta c_{HT}$ correction. Theses leads some implications that the surface turbulence dynamics by wave breaking are interactive with those driven by $\Delta c_{HT}$. In the recent observation work about wind and surface wave dynamics in the coastal ocean (Scully et al., 2015, 2016), they insisted that the surface turbulence by wave breaking can provide the necessary “seed” vorticity to initiate Langmuir turbulence which are
created by the vertical vorticity tilted by Stokes drift. Therefore, the additional observation work is needed to clarify this interaction in western LIS.
Chapter 4. Estimates of surface enhanced turbulent kinetic energy
in a fetch-limited coastal ocean

4.1. Introduction

The turbulent kinetic energy (TKE) in the ocean surface boundary layer largely determines the rates of exchange of momentum and materials between the atmosphere and ocean. It has been long established that the rates of dissipation and production by shear (Arsenyev et al., 1975; Dillon et al., 1981; Oakey and Elliott, 1982; Jones, 1985; Soloviev et al., 1988) are important in the determination of the evolution and structure of the TKE. Recently, observations have demonstrated that the additional generation by surface gravity wave, Langmuir circulation and wave breaking are also important (Kitaigorodskii et al., 1983; Gregg, 1987; Gargett, 1989; Agrawal et al., 1992; Drennan et al., 1996; Anis and Moum, 1992, 1995; Osbron et al., 1992; Drenna et al., 1992b; Melville, 1993, 1994; Craig and Banner, 1995; Greenan et al., 2001; Soloviev and Lukas, 2003; Gemmrich and Farmer, 2004; Stips et al., 2005; Feddersen et al., 2007; Jones and Monismith, 2008b; Gerbi et al., 2008, 2009). Laboratory work (Melville, 1993, 1994; Veron and Melville, 2001) supports this and these mechanisms have been included in modeling studies (Skyllingstad and Denbo, 1995; Mc Williams et al., 1997; Noh et al., 2004; Li et al., 2005; Sullivan et al., 2007; Craig and Banner, 1994; Craig, 1996; Terray et al., 1999b; Burchard, 2001; Umlauf et al., 2003; Kantha and Clayson, 2004).

Measurements by Thompson (2012) of TKE at the ocean surface have shown that wave breaking frequency and the TKE dissipation rate increase as the fetch increases. Henderson et al. (2013) described a parameterization of the effects of wave breaking on the near surface eddy viscosity. Scully et al. (2015) and Scully et al. (2016) presented observations in the surface
boundary layer of the Chesapeake Bay and demonstrated that the interactive effects of breaking waves and Stokes drift shear are important in the deepening of the surface mixed layer. Schwendeman and Thomson (2015) reported observations of whitecap coverage which are the visual signature of wave breaking and showed that the whitecap coverage increases with wind stress and wave slope, and that the near surface turbulent dissipation also increased. Thompson et al. (2016) then demonstrated that five parameterizations of dissipation rate were consistent with the observations.

In this Chapter, I test whether the differences between observations of near surface motion observed in Long Island Sound using HF RADAR (CODAR) and bottom mounted acoustic Doppler current profilers (ADCP) can be explained by the theories that include the effects of near surface intensification of TKE by wave processes.

The setting of the study and the observations are described in section 4.2. In section 4.3, I show that the backscattered acoustic intensity from the ADCPs are related to surface wave breaking. Surface eddy viscosity and dissipation rate are estimated using wave-induced formulae to figure out how much wave breaking turbulence occurs in surface layer in section 4.4 and then in section 4.5, I describe how I estimate the thickness of the Wave Breaking Layer (WBL) and Wave Affected Surface Layer (WASL). Finally, I summarize the results and conclusions in section 4.6.

4.2. Observation setting

I use an extensive set of current, wind and wave measurements from western Long Island Sound that spans the period 2007-2011. Two CODAR (Coastal Ocean Dynamics Application Radar) sites were maintained to measure the near surface flow. They are located at Great Captain
Island (Greenwich, CT) and at Stehli Beach (Bayville, NY). These sites are shown in Figure 1 and the locations are summarized in Table 1. The WLIS buoy was equipped with a sensor to estimate the significant wave height and dominant period, and R.M. Young wind speed and direction, and air temperature and humidity instruments. A bottom mounted ADCP was also located at the site of the WLIS buoy and two others were located to the west at the sites labeled FB02 and EXRK. The locations, acoustic frequency and vertical resolution of the velocity measurements are summarized in Table 1.

Figure 1. Coastal geometry and the locations of the ADCPs (EXRX, FB02 and WLIS) and CODAR sites (G.C. and STLI) in Western Long Island Sound
<table>
<thead>
<tr>
<th>Deployment date</th>
<th>Measurements</th>
<th>Station</th>
<th>Latitude (°N)</th>
<th>Longitude (°W)</th>
<th>Frequency and Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>2007 ~ 2009</td>
<td>ADCP (Acoustic Doppler Current Profiler)</td>
<td>WLIS</td>
<td>40.9541</td>
<td>73.5790</td>
<td>300 kHz with 1m bins</td>
</tr>
<tr>
<td></td>
<td></td>
<td>FB02</td>
<td>40.9253</td>
<td>73.6565</td>
<td>300 kHz with 1m bins</td>
</tr>
<tr>
<td></td>
<td>CODAR (Coastal Ocean Dynamics Application Radar)</td>
<td>Great Captain</td>
<td>40.9820</td>
<td>73.6237</td>
<td>25.3MHz with 1km resolution</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Stehli Beach</td>
<td>40.9087</td>
<td>73.5873</td>
<td>26.19MHz with 1Km resolution</td>
</tr>
<tr>
<td>2010~2011</td>
<td>ADCP (Acoustic Doppler Current Profiler)</td>
<td>WLIS</td>
<td>40.9541</td>
<td>73.5790</td>
<td>300 kHz with 1m bins</td>
</tr>
<tr>
<td></td>
<td></td>
<td>EXRX</td>
<td>40.8781</td>
<td>73.7377</td>
<td>600 kHz with 0.5m bins</td>
</tr>
</tbody>
</table>

Table 1. Details of ADCP and CODAR deployment

4.3. Wind and wave driven surface intensification

The acoustic echo intensity measured in each beam of an ADCPs is related to the backscatter created by suspended particles, (plankton and sediments), density microstructure, bubbles, marine organisms. The backscatter intensity has been used for both the biological scattering to investigate the abundance and distribution of marine organisms (Kaye, 1978; Sund, 1935; Holliday, 1972a; Postel, Lutz, et al., 2007) and energy dissipation signals for surface intensification (Kent, 1999;
Ross and Lueck, 2005; Colbo, Keir, et al., 2014; Demer, David A., et al. 2015). The backscatter intensities from these biological scatters can give the great information how their distribution changes temporally and spatially. Also, the dynamical response of energy dissipation signals are useful to figure out how the energy dissipates depending on the physical factors such as wind and wave variation.

![Figure 2](image)

Figure 2 (a) and (b). Eco intensity of ADCPs at WLIS and at EXRX in 2010 and 2011

Kent (1999) used the sonar equation to relate the backscatter intensity and the echo intensity measured by ADCPs. The instruments report the echo intensity, $I_{\text{counts}}$, and the backscatter (dB) is then
\[ I_{dB} = C \cdot I_{counts} + 20 \log_{10} R + 2\alpha R - 10 \log_{10} \left( \frac{L_{xmit}}{\cos \theta} \right) \]  

(1)

where \( R = \frac{r^{0.5} L_{xmit}}{\cos \theta} \), \( r \) is the range from the transducer to the middle of the bin, \( \theta \) is beam angle from vertical, \( L_{xmit} \) is the transmit length, \( \alpha \) is sound absorption coefficient and \( C \) is echo intensity coefficient, \( C = 127.3/(T_e + 273) \), where \( T_e \) is the temperature of the transducer (°C).

Figures 2 show the evolution of the vertical structure of the echo intensity, \( I_{counts} \), at WLIS and EXRK between 2010 and 2011. Both records clearly indicate that the echo intensity is intensified near the surface layer (~6m). Figure 3 shows the echo intensity during the passage of two hurricanes, Hurricane Earl (2010) and Hurricane Irene (2011). The surface set-up and the increased amplitude of the surface backscatter evidently vary as the hurricane passes through the region.

To investigate the relationship between the surface backscatter and the significant wave height we used Equation (1) to convert the echo intensity to the backscatter and then normalized the backscatter intensity in the surface bin with the value in the bottom bin. Figure 4 shows the dependence of the normalized surface backscatter intensity on the significant wave height. The blue lines represent bin averages of all the data years while the red indicates averages of the data acquired during the hurricane passages shown in Figure 3. These results provide clear evidence that the surface intensified signals are strongly dependent on the wind and wave in the ocean surface layer.
Figures 2 and 3 clearly indicate that echo intensity increases in the surface layer both in the total aggregated years and in the hurricanes. There are two hurricanes affecting largely in Long Island Sound in 2010 and 2011. Hurricane Earl (Category 1 hurricane) was formed on August 25th and dissipated on September 6th in 2010. This had affected in Long Island Sound from August 30th to September 6th. In western LIS during Hurricane Earl, the highest wind speed is about ~0.2 Pa and highest significant wave height is about 0.7m. Hurricane Irene (Category 3 hurricane) on August 21th and dissipated on August 30th in 2011. This had affected in Long Island Sound from August
28th to August 30th. In western LIS during hurricane Irene, the highest wind speed is about ~0.4 Pa and highest significant wave height is about 2.2 m.

Figure 4 (a). Bin averaged, normalized surface backscatter intensity at WLIS dependence on significant wave height. The blue line shows the average of all data, and the red line show the data from the Hurricane intervals. 4(b) shows the same analysis at EXRX.

When wind and wave energy are strong during hurricane events, the surface intensified signals are more enhanced than the normal time as shown in Figure 4 such that the normalized backscatter intensity increases with significant wave height. Additionally, the difference between blue and red lines in Figure 4 shows how the surface turbulence is dependent on the occurrence of high waves. As the several studies (Stoker, 1957; LeMehaute, 1962; Battijes and Jansen, 1978; Battjes and Stive, 1985) showed that the occurrence of wave breaking is important to consider to
figure out the dissipation energy on the surface, the difference of red and blue lines indicate the difference of wave breaking occurrence between the normal time and hurricanes. During hurricanes, the more occurrence of higher wind results in the more occurrence of higher wave so that the normalized backscatter intensity dependence on the wave height increases. This is because the surface turbulence energy dissipation through the wave breaking is intensified more when the wave field is young in the extreme wind events during hurricanes.

Figure 5. Wave steepness in the function of along-Sound wind stress.

To determine how much surface wave breaking occurs in western LIS, the wave steepness (S) is calculated as
\[ S = 0.5 \frac{H_s}{\lambda} \]  \hspace{1cm} (2)

Where \( H_s \) is the significant wave height, \( \lambda \) is the wave length estimating using measured wave period. Figure 5 shows the wave steepness increases in the function of wind stress. This indicates the wave breaking occurs more as wind increases. It is common that whitecap coverage indicates the surface wave breaking and the resent observation study (Schwendeman et al., 2015) shows that the whitecap coverage over the ocean surface increases rapidly when the wave steepness is exceeds 0.01. Their results also show the whitecap coverage is significantly large enough for the significant wave breaking at \(~0.013\). Adapting this criterion (~0.013), Figure 5 suggests for western Long Island Sound that the threshold to appear the wave breaking is reached when the along-Sound wind stress exceeds \(~0.1\) Pa.

4.4. Estimates of surface intensified dissipation rate and eddy viscosity

In order to understand the wind and wave induced surface turbulence, both dissipation rate and eddy viscosity are good indicators of how much turbulence transfers from the atmosphere to the ocean surface. Energy dissipation \( \varepsilon \), which is the loss rate of the turbulent kinetic energy per unit mass through viscosity to heat, is generally used as a measure to characterize ocean turbulence. Eddy viscosity \( \nu \) represents the quantity of momentum transfer which is formulated with the mean velocity gradient (shear) \( \frac{du}{dz} \) and momentum flux \( \tau \) (Thorpe, 2005).

Henderson et al. (2013) suggests that using the wave age parameter, significant wave height and friction velocity the wave-induced surface eddy viscosity (\( \nu \)) and the dissipation rate(\( \varepsilon \)) are

\[ \nu = \alpha_\nu u_* H_s \]  \hspace{1cm} (3)
\[ \varepsilon = \frac{\alpha_v u_*^3}{H_s} \] (4)

Where \( \alpha_v \) is wave age related empirical constant, \( u_* \) is water-side friction velocity and \( H_s \) is the significant wave height. The wave age related empirical constant \( \alpha_v \) is expressed as

\[ \alpha_v = k \cdot w_a^{\frac{1}{3}} \] (5)

where \( w_a \) is wave age \((c/u_{air})\) and \( k \) is the empirical constant.

Using Equation (5), Equation (3) is converted to

\[ \nu = k \cdot w_a^{\frac{1}{3}} \cdot u_* \cdot H_s \] (6)

There is a general way to estimate eddy viscosity so called simple gradient transfer method

\[ \nu_s = \frac{\tau}{(\rho \frac{du}{dz})} \] (7)

Assuming that the wave-induced \( \nu \) is equal to \( \nu_s \) with simple gradient transfer method, \( k \) is derived.

\[ k = \frac{u_*}{w_a^{\frac{1}{3}} \cdot H_s \cdot \frac{du}{dz}} \] (8)

Using the relationship between the dissipation rate and eddy viscosity which is from Thorpe (2005) for the case of isotropic turbulence,

\[ \varepsilon = \frac{15}{2} \cdot \nu \cdot \frac{\tau}{\rho_0} \] (9)

The empirical constant \( k \) is also expressed with \( \alpha_v \).
\[ k = \frac{\left( \frac{2}{15} \right) \alpha \cdot u_s^2}{w_a^2 \cdot (H_s \cdot \frac{du}{dz})^2} \]  \hspace{1cm} (10)

Where \( \alpha_e = \alpha \cdot w_a \) and \( \alpha \) is wave age related parameter (Terray et al., 1996). Using the measured surface shears \( \frac{du}{dz} \), empirical constant \( k \) and significant wave height \( H_s \), the wave-induced eddy viscosity and dissipation rate are estimated in the following

\[ \nu = k \cdot w_a^\frac{1}{3} \cdot u_s \cdot H_s \]

\[ = \frac{w_a \cdot \alpha \cdot u_s^3}{H_s \cdot \frac{du^2}{dz}} \]  \hspace{1cm} (11)

As shown in Terray et al. (1996) the \( \alpha \) is effective phase speed forced by wind relative to the wave phase speed and increases as the wave age decreases. Generally, it is generally about 0.5 in the young open ocean and it is assumed to be about 0.3 in the shallow coastal water. While there have been many studies (Hasselmann et al., 1973; Kahma, 1981; Donelen et al., 1985; Birch and Ewing, 1986 and Terray et al., 1996) to figure out the relation between \( \alpha \) and wave age for the open ocean condition, it is not evident with any observation studies yet for the coastal ocean. Therefore, eddy viscosity and dissipation estimation along with the \( \alpha \) parameterization should be essential to clarify the wave induced surface dynamics in shallow coastal water.

Figure 6 shows the surface shears in response to along-Sound wind after taking out the Stokes drift (Ch. 3). With the surface shears in Figure 6, the estimated eddy viscosity and dissipation rate using wave-induced method are shown in Figure 7 (a) and (b) with three different \( \alpha \) in response to along-Sound using Equation (9) and (11). Both eddy viscosity and dissipation rate are asymmetric that they are larger in the negative along-Sound wind while they are smaller in the positive along-Sound wind.
Figure 6. The averaged surface shears in the WASL in response to along-Sound wind.
Figure 7. Estimation of Eddy viscosity and Dissipation rate in WASL in response to along-sound wind to compare using the simple gradient transfer method and the wave induced method with three different wave age related parameter $\alpha$. 
This indicates that the surface mixing and turbulent kinetic energy transfer through wave breaking are highly dependent on the asymmetric wave dynamics (Ch. 2) in western Long Island Sound. Figure 7 shows that the eddy viscosity and dissipation rate increases as $\alpha$ increases.

Furthermore, I found wave age parameter $\alpha$ is 0.13 in negative along-Sound wind while it is 0.2 in the positive along-Sound wind in western Long Island Sound which results from the comparison of observation and three different cases with different $\alpha$ shown in Figure 8.

![Figure 8](image_url)

Figure 8. Comparison of observational surface shears (Blue) and the estimated surface shears with constant eddy viscosity (grey dash line), the wave-induced eddy viscosity with $\alpha$ is 0.1 (red line), 0.13 (red dashed line), 0.2 (magenta line) and 0.3 (black line). The blue dashed line represents the linear regressed line for the observation.
In Figure 8, the observed surface shear (blue line) is compared with the estimated surface shear with different $\alpha$ and they are most consistent when $\alpha$ is $\sim 0.13$ in the negative along-Sound wind and it is $\sim 0.2$ in the positive along-Sound wind in western Long Island Sound rather than the higher values which are used in deep ocean. Comparing with the constant eddy viscosity in grey dashed line to the observed results, they are not consistent each other. Therefore, I found out the eddy viscosity increased in the function of wind stress asymmetrically.

4.5. Estimates of wave-induced surface intensification

Wave-affected surface layer ($Z_w$) and wave-breaking layer ($Z_b$) thickness determined by wave age and significant wave height indicate how deep turbulent kinetic energy is distributed in the surface boundary layer. The energy source driving surface flow is transferred from the wind to the water column by the waves and an estimation of wind energy input to the waves provides an approximate indication of the energy input generating turbulence and physical forcing in the near surface (Terray et al., 1996; Gerbi et al., 2009; Thompson et al., 2015). To estimate wind energy input, the simple related formula $E_0 = -u_*^2 G_e$ from previous studies (Craig and Banner, 1994; Terray et al., 1996; Sullivan et al., 2007; Gerbi et al., 2009) is used where $u_*$ ($= \frac{\tau}{\sqrt{\rho}}$) is water-side friction velocity and $G_e$ is an empirical function of the wave age. The “effective phase speed” is defined as $c$ relating to wind input which is $E_0 (\equiv \frac{\tau_* c}{\rho_w} \approx u_*^2 c)$. 

With the observed wave and wind input estimation, Terray et al. (1996) showed that wave-breaking layer thickness ($Z_b$) is

$$\frac{Z_b}{H_s} \approx 0.6 \tag{12}$$
Terray et al. (1996) shows the thickness of WASL calculating with $\alpha$ as

$$\frac{Z_w}{H_s} = 0.3 \kappa \alpha c_p / u_*$$

(13)

Where $\kappa$ is Von Kaman constant (~0.4), $\alpha$ is the wave age parameter, $c_p$ is the phase speed and $u_*$ is the friction velocity.

Using Equation (12) and (13) with the measured significant wave height, phase speed, estimated $\alpha$ in the previous section, wave breaking layer thickness ($Z_b$) and wave affected surface layer thickness ($Z_w$) are presented for positive and negative along-Sound wind in different wind stress magnitude in Table 2.

<table>
<thead>
<tr>
<th></th>
<th>Negative $\tau_x$</th>
<th>Positive $\tau_x$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1Pa</td>
<td>0.75m</td>
<td>0.45m</td>
</tr>
<tr>
<td></td>
<td>0.45m</td>
<td>0.27m</td>
</tr>
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<td></td>
<td>5.8m</td>
<td>4.6m</td>
</tr>
<tr>
<td>0.2Pa</td>
<td>1.1m</td>
<td>0.7m</td>
</tr>
<tr>
<td></td>
<td>0.66m</td>
<td>0.42m</td>
</tr>
<tr>
<td></td>
<td>6.8m</td>
<td>5m</td>
</tr>
<tr>
<td>0.3Pa</td>
<td>1.55m</td>
<td>0.9m</td>
</tr>
<tr>
<td></td>
<td>0.93m</td>
<td>0.54m</td>
</tr>
<tr>
<td></td>
<td>8.6m</td>
<td>6m</td>
</tr>
</tbody>
</table>

Table 2. Measured significant wave height (Black), the estimated WBL (Blue) and WASL (red) in response to different wind stress

Because of the larger wave in the negative along-Sound wind, the thickness of WBL and WASL is deeper in the negative along-Sound wind than in the positive along-Sound wind as shown in Table 2. The average depths of WBL and WASL are 0.68m and 6.6m in the negative along-Sound wind while 0.41m and 5m in the positive along-Sound wind. This asymmetry of thickness indicates that the TKE dissipation enhancement with wave breaking penetrates more deeply in negative along-Sound wind rather than positive along-Sound wind in western LIS.
4.6. Summary and Conclusions

Surface wave breaking plays a major role in explaining the surface energy balance generating strong turbulent motion and energy dissipation. This wave breaking limits wave growth, induces more air-sea gas exchange and the turbulence that generates surface ocean mixing (Thomson et al. 2016). The strong surface turbulence by wave breaking generated by wind controls how deep and how much the vertical mixing in the surface layer occurs through dissipating turbulent kinetic energy and transferring turbulent momentum flux.

In this chapter, the measured surface turbulent signals and wave steepness are estimated with the observation data in wind and wave functions in order to figure out how wind and surface waves can produce ocean surface turbulence. The formulation to estimate WASL (Wave Affected Surface Layer) and WBL (Wave breaking layer) thickness are shown to define the depth affected by surface turbulence. The wave-induced dissipation rate and eddy viscosity are estimated using Henderson’s parameterization equations to explain how much turbulent energy dissipates in the estimated WASL.

First, the surface intensified backscatter signals with ADCPs are shown as the evidence of surface turbulence induced by wind and wave both in normal time and hurricane time. As expected, the surface intensified signals are significantly larger and clearer and the backscatter intensity is increased more with significant wave height during hurricanes than in normal time. Otherwise, when the wind and wave energy is extremely strong enough to generate strong turbulence, the dissipation of turbulent energy and surface mixing occurs more. This is the apparent evidence that the wind and wave are the main factors to induce the surface ocean turbulence. Then, wave steepness calculated to show how much waves break in the function of wind. With the wave
breaking criteria (Schwendeman et al., 2015), the wave breaking occurs significantly when the wind stress is larger than ~ 0.1 Pa.

Additionally, the strong diurnal periodic backscatter signals are found vertically during Hurricanes (Figure 3). This indicates the zooplankton migration diurnally. As several studies (Postel, Lutz, et al., 2007; Lebourges-Dhaussy, Anne, et al., 2009; Skjoldal, Hein Rune, et al., 2013) show that backscatter intensity can be used to figure out the distribution of zooplankton diurnal vertical migration and to estimate the biomass of them, these signals during hurricanes are interesting to figure out the marine organism movements driven by physical energy particularly when the high turbulent energy is transferred from the surface because of hurricanes.

Second, the eddy viscosity and dissipation rate are estimated to figure out how the turbulent momentum flux and turbulent kinetic energy is transferred in WASL. To estimate the surface wave-induced eddy viscosity and dissipation rate, Henderson’s parameterization equations are used taking into account surface wave breaking. The empirical constant k is estimated with the measured data and the three different wave age parameters ($\alpha$) of 0.1, 0.13, 0.2 and 0.3 are used. 0.13 appears to be the most reliable value in the negative along-Sound wind and the model with 0.2 is best in the positive along-Sound wind for western Long Island Sound. This is because the wave age is younger when the wind is positive along-Sound wind than in the negative along-Sound wind. Asymmetry is also found in the estimated eddy viscosity and dissipation rate and this asymmetry of eddy viscosity and dissipation rate indicates that the near surface wave-induced eddy viscosity and dissipation rate result from the asymmetric currents and shears on the ocean surface. So, the different wind-fetches generate the wave differently to produce turbulence kinetic energy and momentum flux asymmetrically.
Lastly, WASL and WBL thickness is estimated using observation data and parameterized $\alpha$. Terray et al. (1996) showed the equations to estimate WBL and WASL thickness using $\alpha$ and the equation is modified with the parameterized $\alpha$ for western LIS case. The averaged WASL thickness is 6.6 m in negative along-sound wind and 5.2 m in positive along-sound wind and the averaged WBL thickness is 0.66 m in negative along-sound wind and 0.4 m in positive along-sound wind. This indicates that the depth at which the surface turbulence dissipates is asymmetric because of the asymmetry of the wind and wave dynamic response to along-Sound wind.

Therefore, I conclude that the surface TKE is asymmetrically dissipated in the surface layer caused by the asymmetric wave dynamics in western LIS. This leads us to figure out how the turbulent mixing through surface wave breaking on the surface layer is in response to wind and wave characteristics in western LIS.
Chapter 5. Modeling Wind and Wave Driven circulation in a fetch-limited coastal environment:
western Long Island Sound

5.1. Introduction

The circulation in semi-enclosed estuarine basins and coastal bays is largely considered by the wind, tide, buoyancy forces and the Earth’s rotation. There have been several studies showing the estuarine circulation affected by wind (Weisberg, 1976; Elliott, 1978; Wang, 1979; Noble et al., 1996; Geyer, 1997; Scully et al., 2005; Chen and Sanford, 2009; Whitney and Codiga, 2011). Geyer (1997), Scully (2005) and O’Donnell et al. (2014) show how wind straining and mixing interacts with the estuarine circulation by buoyancy in the observation and Li and Li et al. (2012) describes the model results of the interaction between wind and buoyancy driven estuarine circulation.

Pritchard (1967), Friedrichs and Hamrick (1996), Hendershott and Rizzoli (1976) describe how the earth’s rotation is essential in explaining circulation in coastal, estuarine and lake water. Kasai et al. (2000) and Valle-Levinson et al. (2003) describe the effect of Coriolis force on buoyancy-driven flows with Ekman solution. Winant (2004) show the analytical solution of the wind-driven barotropic model to take into account wind force, pressure gradient force and Coriolis force for the coastal semienclosed and elongated basins. Whitney and Codiga (2011) also show the compared results of numerical simulation (Regional Ocean Modeling System) and analytical solutions (Csanady, 1973; Winant, 2004) taking into account the stratification and rotation effect with the Ferry-based observation data in eastern Long Island Sound. Their modeling results along with the observation results in eastern LIS show the asymmetric response to along-Sound wind,
i.e. there is more stratification in westward wind while there is more mixing in eastward wind in eastern LIS because the Connecticut river input is westward in eastern LIS. In western LIS, in contrast, their model results showed there is more mixing in westward wind while there is more stratification in eastward wind. These results indicate that the fresh water effect interacting with the wind driven circulation is important to consider the water circulation in the coastal water.

However, none of these studies considers the effect of surface gravity waves on the circulation of water and observation analyses and model studies in the coastal and estuarine ocean have not included the effect of surface wave although the surface wave energy by wind is an important part of water circulation not only on the surface but also in the entire water column. The main goal of this work is to estimate how turbulent energy production by breaking surface waves affects water circulation in a coastal basin.

**5.1.1. Turbulent Kinetic Energy Balance**

Wind stress is complicated to the estuary water by turbulence. The Turbulent Kinetic Energy (TKE) equation is written as (Edson and Fairall, 1998)

\[
-u'w' \frac{\partial \bar{u}}{\partial z} - v'w' \frac{\partial \bar{v}}{\partial z} + g \left( \frac{\partial \bar{w}' \theta'}{\partial z} \right) - \frac{\partial \bar{w}' \rho'}{\partial z} - \frac{1}{\rho} \frac{\partial \bar{p}'}{\partial z} - \varepsilon = 0
\]

(1)

where \( \bar{u}, \bar{v} \) and \( \bar{w} \) are the Reynolds-averaged velocity vector components, \( u', v' \) and \( w' \) are the turbulent perturbations, \( \bar{\rho} \) is the mean water density, \( p' \) is the pressure perturbation, the mean kinetic energy \( \bar{e} = \frac{1}{2} (\bar{u}'^2 + \bar{v}'^2 + \bar{w}'^2) \), \( \bar{\theta}' \) is the mean virtual potential temperature and \( g \) is the gravitational acceleration. The first two terms are mechanical production terms, the third is a buoyant production and consumption term, the fourth is a turbulent energy transport term, the fifth is a pressure gradient term and the last term represents the dissipation rate.
The mechanical production terms represent turbulence production due to vertical shear in the horizontal flow. The buoyant production and consumption term indicates the buoyancy flux by stability. When this term is positive, TKE is produced due to the thermal energy. When this term is negative in the stable stratification, the TKE is transferred to the potential energy. The turbulent energy transport term represents TKE flux divergence. This term explains how the kinetic energy is transported to the ocean and generally involves the loss of TKE in the surface boundary layer. Pressure flux divergence describes how TKE is redistributed by pressure perturbations. In the surface boundary layer, this term becomes to gain TKE which compensates for the loss from the energy transport term. Turbulence decays into viscous motion by the molecular processes so that TKE dissipates into heat energy.

5.1.2. Surface enhanced turbulent energy by wave breaking

When the amount of energy transport (loss) is equal to the amount of pressure transport (gain) and the buoyancy term is negligible in the neutral condition, the dissipation is nearly in balance with the mechanical production term in the surface boundary layer. This is the classical boundary theory known as log law. However, the surface intensification of turbulent energy by wave breaking is significant enough to modify the turbulent energy balance in the ocean boundary layer such that the kinetic energy transport from breaking waves induced by wind energy input is enhanced. So, if the wave breaking is taken into account, the dissipation rate is more balanced with the kinetic energy transport rather than mechanical shear production resulting in a dependence from the log log profile shown in Figure 1 from Gerbi et al. (2009).
Several field studies (Kitaigorodskii et al., 1983; Gregg, 1987; Gargett, 1989; Agrawal et al., 1992; Drennan et al., 1996; Anis and Moum, 1992, 1995; Drennan et al., 1996; Melville, 1996; Craig and Banner, 1995; Greenan et al., 2001; Soloviev and Lukas, 2003; Gemmrich and Farmer, 2004; Stips et al., 2005; Feddersen et al., 2007; Jones and Monismith, 2008b; Gerbi et al., 2008, 2009), modeling studies (Skyllingstad and Denbo, 1995; Mc Williams et al., 1997; Noh et al., 2004; Li et al., 2005; Sullivan et al., 2007; Craig and Banner, 1994; Craig, 1996; Terray et al., 1999b; Burchard, 2001; Umlauf et al., 2003; Kantha and Clayson, 2004) and laboratory work (Melville, 1993,1994; Veron and Melville, 2001) have shown clearly that surface intensified turbulence such as that produced by surface gravity waves, Langmuir turbulence and wave breaking should be considered to explain circulation. Henderson et al. (2013) showed that the surface enhanced mixing induced by wave breaking with observational shear and the parametric eddy viscosity model in the shallow tidal flat. Scully et al. (2015) and Scully et al. (2016) show the observation
results that the vertical mixing in the surface boundary layer is efficient in wave-driven turbulence for the shallow coastal water such as Chesapeake Bay. They demonstrate the wave-driven turbulence results from the interactive effects of breaking waves and Stokes shears. They conclude that breaking waves by wind energy input initiate the surface intensification and then the near surface mixed layer deepens with Stokes shear. Schwendeman and Thomson (2015) describes the observation results of whitecap coverage which are the visual signature of wave breaking. They provide the observation evidence that the whitecap coverage increases with wind stress, wave slope and turbulent dissipation increase. Thompson et al. (2016) demonstrated that the analytical formulations of dissipation rate are consistent with the observation results of kinetic energy by surface wave breaking.

In this chapter, the wind-driven barotropic model of Winant (2004) is modified to include the effect of surface enhanced turbulence by wave breaking and is compared with the observation data in a fetch-limited coastal basin, western Long Island Sound. In the following Section 5.2, I show the Winant (2004) wind-driven barotropic model equations to produce Wave and wind driven barotropic semi-Analytical Model with depth-dependent eddy viscosity. In Section 5.3, the observational setting and the observation results of mean circulation in response to wind are shown for a series of sites in LIS. In Section 5.4, three different depth-dependent eddy viscosity models are presented and evaluated according to the different boundary theories, i.e. rigid boundary theory, wave surface boundary theory and those without boundary dynamics. In Section 5.5, the model results with three different eddy viscosity models are compared with the observation data. Then, these are summarized in Section 5.6.
5.2. Wave and wind driven barotropic semi-Analytical Model

5.2.1. Wind-driven momentum equation with depth-dependent eddy viscosity

The wind-driven barotropic steady state $x$ and $y$ momentum equation such that the Coriolis force is in balance with pressure gradient and friction force is

$$
-f \cdot v = -\frac{1}{\rho} \frac{\partial P}{\partial x} + \frac{1}{\rho} \frac{\partial \tau}{\partial z}
$$

(2)

$$
= -\frac{1}{\rho} \frac{\partial P}{\partial x} + K_z \frac{\partial^2 u}{\partial z^2}
$$

$$
f \cdot u = -\frac{1}{\rho} \frac{\partial P}{\partial y} + \frac{1}{\rho} \frac{\partial \tau}{\partial z}
$$

(3)

$$
= -\frac{1}{\rho} \frac{\partial P}{\partial y} + K_z \frac{\partial^2 v}{\partial z^2}
$$

where $P$ is the hydrostatic pressure, $\tau$ is surface stress and $\rho$ is water density, $f$ is the Coriolis parameter and $K_z$ is the eddy viscosity.

The left side of the equations is the Coriolis acceleration term, the first term in the right side is pressure gradient and the second term is friction force with constant eddy viscosity. In order to figure out the dynamics with a depth-dependent eddy viscosity, the momentum equation should be modified to
\[-f \cdot v = -\frac{1}{\rho} \frac{\partial P}{\partial x} + \frac{1}{\rho} \frac{\partial \tau}{\partial z} \]  
\(= -\frac{1}{\rho} \frac{\partial P}{\partial x} + \left( \frac{\partial K_z}{\partial z} \frac{\partial u}{\partial z} + K_z \frac{\partial^2 u}{\partial z^2} \right) \tag{4}\)

\[f \cdot u = -\frac{1}{\rho} \frac{\partial P}{\partial y} + \frac{1}{\rho} \frac{\partial \tau}{\partial z} \]  
\(= -\frac{1}{\rho} \frac{\partial P}{\partial y} + \left( \frac{\partial K_z}{\partial z} \frac{\partial v}{\partial z} + K_z \frac{\partial^2 v}{\partial z^2} \right) \tag{5}\)

Following Winant (2004), the complex form of Equations (4) and (5) is

\[fk \times u^* = -g \nabla \cdot \eta^* + K_z \frac{\partial^2 u^*}{\partial z^*} + \frac{\partial K_z}{\partial z^*} \frac{\partial u^*}{\partial z^*} \]  
\tag{6}\)

To make a non-dimensional equation, the relationship between dimensional (starred) variables and non-dimensional (unstarrd) variables becomes

\[(x, y, \ell) = \frac{x^*, y^*, \ell^*}{B^*}, z = \frac{z^*}{h_0^*}, h = \frac{h^*}{h_0^*} \tag{7}\)

\[u = \frac{\rho Ku^*}{\tau_s h_0^*}, w = \frac{\rho B^* Kw^*}{\tau_s h_0^*} \tag{8}\)
\[
\eta = \frac{\rho g h_0 \eta^*}{|\tau_s| B^*}
\]  

(9)

where 2L* is length, 2B* is basin width, h₀ is maximum depth and \( \tau_s \) is surface wind stress. Figure 2 shows basin geometry used in the model.

Figure 2. The geometry of a simple basin of length 2L and Width 2W (Winant 2004)

Using the relation between dimensional and non-dimesional variables in Equations (7), (8) and (9), the non-dimensional version of Equation (6) can be converted to the Equation (10).

\[
\frac{2}{\delta^2} k \times u = -\nabla \eta + \frac{1}{K_z} \frac{\partial K_z}{\partial z} \frac{\partial u}{\partial z} + \frac{\partial^2 u}{\partial z^2}
\]  

(10)

where Ekman depth \( D_E = \frac{2K}{\sqrt{f}} \) and the non-dimensional Ekman depth \( \delta = \frac{2k}{\sqrt{f h_0^2}} \).
The delta $\delta (= \frac{D_E}{h_0})$ is non-dimensional Ekman depth, which is a ratio of Ekman depth ($D_E$) and maximum depth ($h_0$). Winant (2004) calls the inverse delta $\delta^{-1} (= \frac{h_0}{D_E})$ as the rotation factor because the rotation factor relative to the friction effect is more as the inverse delta increases. For depth-dependent eddy viscosity cases, delta is maximum value of it.

Defining the complex variable and coefficients as velocity $U = u + iv$ and the pressure gradient $N = \eta_x + i\eta_y$, the momentum equation is finally rewritten as

$$\frac{d^2U}{dz^2} + \frac{1}{K_z} \frac{dK_z}{dz} \frac{dU}{dz} - \frac{2}{\delta^2} U = N$$

(11)

This momentum equation is semi-analytically solved for being in the middle of the basin. Since the axial gradients are negligible in the middle of the basin and the lateral and vertical variation is more significant.

5.2.2. Semi-analytical solution

Winant (2004) showed that for the constant $K_z$ could be solved analytically in a semi-closed elongate basin. This analytical solution is described in Appendix A.

The transport stream functions and the vertically integrated form are described for the middle of basin from Winant (2004) as

$$\psi_x = \int_{-h}^{0} v \, dz \quad \text{and} \quad \psi_y = -\int_{-h}^{0} u \, dz$$

(12)
\[ 2 \delta^{-2} \nabla \psi = h \nabla \eta - t_s + T_b \quad (13) \]

The boundary condition in mid-basin is that the lateral transport is zero for all \( y \) locations. So, the pressure gradient and bottom stress are

\[ h \eta_x + T_b^x = t_s^x \quad (14) \]

\[ h \delta \eta_y - \delta t_s^y + \delta T_b^y = 2 \psi_s^y \quad (15) \]

where \( T_b^x \) and \( T_b^y \) are bottom stress, \( t_s^x \) and \( t_s^y \) are surface wind stress in \( x \) and \( y \) directions.

Figure 3. (a) shows the model results comparison of analytical solution and numerical solution such as finite element method (FEM) and finite difference method (FDM) and (b) shows how \( e_n^{xy} \) is converged and \( N \) is the number of iteration.
To solve momentum Equation (11) with the depth-dependent eddy viscosity, the iteration method is established with iteration error function using Equations (14) and (15) in x and y direction are

\[ e_n^x = h\eta_x + T_b^x - t_s^x \]  \hspace{1cm} (16)

\[ e_n^y = h\delta\eta_y - \delta t_s^y + \delta T_b^y - 2\psi_y \]  \hspace{1cm} (17)

And the magnitude of error function is

\[ e_n^{xy} = \sqrt{(e_n^x)^2 + (e_n^y)^2} \]  \hspace{1cm} (18)

In order to calculate this iteration error function, there are several steps. First, Equation (11) is solved with the prescribed bottom stress and pressure gradient which are from Winant’s analytical solution with the constant eddy viscosity and in the second step, the bottom stress and pressure gradient are calculated with the u and v from the first step using Equation (14) and (15). Prescribing the depth-dependent eddy viscosity, the bottom stress and pressure gradient estimated from the second step, Equation (11) is solved again while I calculated the iteration error function with the new pressure gradient, bottom stress, u and v using Equation (16) and (17). These processes are iterated until the iteration error function is converged to the minimum values (Figure 3 (b)) to find out the most accurate solution.
Figure 3 (a) and (b) show the performance and accuracy of numerical model (Figure 3(a)) and convergence of the iteration processes (Figure 3(b)) to find the best solution for the momentum equation with depth-dependent eddy viscosity using the iteration error function $e_{n}^{xy}$. The model runs with one of the Finite element methods such as Galerkin’s method and the boundary condition is solved with a shooting method because Figure 3 (a) shows the analytical model results (Winant 2004) are more consistent with Finite element method (Galerkin’s method) rather than finite difference method (Euler’s method). RMSE (Root Mean Square Error) between analytical solution and finite element method is about $4 \times 10^{-4}$ while it is $3 \times 10^{-3}$ between analytical solution and finite difference method. The resolution of the numerical method is 0.1m. The details of the selected numerical solution (Galerkin’s method) are described in the Appendix B.

5.3. Observation setting and wind-driven currents

The western end of Long Island Sound is a narrow elongate basin resulting in fetch-limited wave dynamics which affect the circulation and mixing of the water column depending on the direction of the wind. When the wind is from east, waves are larger due to longer fetch while the wave is smaller with a shorter fetch if the wind is from west (See Chapters 2 and 3).
To examine the response of the circulation field to winds and waves, I use two observation instruments located in mid-basin. The first instrument is CODAR (Coastal Ocean Dynamics Application Radar) which measures near surface currents (~5m) and the second instrument is the ADCP (Acoustic Doppler Current Profiler) to measure currents in the rest of the water column. The ADCPs are located at the stations in mid-basin (WLIS and FB02 in Figure 4) and CODARs are sited one on Great Captain (G.C.) screening the Connecticut side and the second at STLI (Stehli) on the Long Island side. CODAR radial surface currents are collected near the ADCP locations and are used to calculate total current velocity. Both the resulting CODAR and ADCP total current velocity is rotated to take the along-sound and across-sound current velocity with rotation matrix.
The ADCP were almost co-located with a surface buoy providing wave heights and wind speed and direction (The distant between ADCP and buoy is about 0.2 km). As described in the previous chapters, the asymmetric wave dynamics in the along-Sound regime results in an asymmetry of significant wave height, surface currents, surface shears, surface eddy viscosity and dissipation rate.

Figure 5. Vertical structure of wind-driven circulation in the along-sound wind. The u and τₓ represent along-sound velocity and along-sound wind stress. Blue indicates the negative wind and red represents the positive wind in each wind event. The solid line indicates the vertical profiles using CODAR data after taking out surface Stokes drift and the dashed line indicate those before taking out surface Stokes drift.
Using the surface data with CODAR and subsurface data with ADCP, Figure 5 shows wind-driven circulation in response to positive (Red) and negative along-sound (Blue). The observation data is refined in the several steps. First, since across-Sound wind correlates the along-Sound wind, the wind-driven currents are sorted out to reduce the correlation in the orthogonal component of the winds (Ch. 2). Then, the surface Stokes drift (Chavanne (2018)) are taken out from the CODAR surface data (Ch. 3) because ADCP can’t measure the Lagrangian Stokes drift while CODAR surface data can measure it (Teague (1986); Fernandez et al. (1996); Paduan and Rosenfeld (1996); Graber et al. (1997); Laws (2001); Ullman et al. (2006); Mao and Heron (2008); Ardhuin et al. (2009); Kirincich et al. (2012)). The dashed line indicates the vertical structure of currents including surface Stokes drift and the sold line represents those after taking out surface Stokes drift in Figure 5. The surface currents structures are asymmetric both in dashed line and solid line and this asymmetry increases in the dashed line after taking out the surface Stokes drift from CODAR data. Both along-Sound currents in the positive and negative along-Sound wind are downwind while there appears upwind current below 4~5m of water depth from the surface.

5.4. Eddy Viscosity Models

I now compare the consequences of eddy viscosity structures, i.e. rigid boundary, surface wave boundary and those without boundary dynamics. First, eddy viscosity is constant in the water column such that the mixing rate is the same over the vertical and is independent of vertical distribution of seasonal and estuarine circulation varied with temperature and salinity. Second, eddy viscosity is structured with the log law theory stating that the logarithm of the distance from the boundary is proportional to the average turbulent velocity of the water.
Lastly, the surface boundary is not rigid and is free and fluctuates between atmosphere and ocean in the realistic ocean surface. So, the eddy viscosity is created by the wave induced surface boundary theory such that the strong enhancement in the ocean surface occurs with the values that are much greater than those predicted by rigid boundary theory. Several research studies have shown that the wave-induced surface boundary theory is more consistent with the realistic ocean (Kitaigorodskii et al., 1983; Gregg, 1987; Gargett, 1989; Agrawal et al., 1992; Drennan et al., 1996; Anis and Moum, 1992, 1995; Drennan et al., 1996; Melville, 1996; Craig and Banner, 1995; Greenan et al., 2001; Soloviev and Lukas, 2003; Gemmrich and Farmer, 2004; Stips et al., 2005; Feddersen et al., 2007; Jones and Monismith, 2008b; Gerbi et al., 2008, 2009).

In the rigid boundary theory, the dissipation rate of the near surface is mostly in balance with shear production which is classically known and consistent with the log law theory. On the other hand, a lot of studies have found that the near surface dissipation rate is in balance with the kinetic energy transport rather than shear production because of the wave breaking in wave induced surface theory (Terray et al., 1996; Gerbi et al., 2009; Henderson et al., 2013; Thompson et al., 2016).

Henderson et al. (2013) showed the parametric equations of depth-dependent drag coefficient, eddy viscosity and dissipation rate taking into account wave breaking in shallow surface water. Their parameterization known as ‘Henderson’s parameterization’ is used and summarized in the following to compare it with rigid boundary theory. In the theory of rigid boundary model, the velocity is $u(z) = \Delta u = \frac{u^*}{k} \log \left( \frac{z_2}{z_1} \right)$ and the boundary stress is $\tau = \rho C_d \Delta u^2$, the drag coefficient is $\sqrt{C_D} = \left. \frac{k}{\log \left( \frac{z_2}{z_1} \right)} \right| = \frac{u_*}{\Delta u}$, the eddy viscosity is $\nu = -u^2 \left( \frac{\partial u}{\partial z} \right) = \kappa u z$ and the
dissipation rate of kinetic energy is \( \varepsilon = \frac{u^3}{\kappa z} = v \left( \frac{\partial u}{\partial z} \right)^2 \) where \( \Delta u is u(z_2) - u(z_1) \), \( \kappa \) is von-kámen constant (~0.4) and \( u_* (u_* = \sqrt{\tau/\rho}) \) is a friction velocity.

In wave affected surface theory taking into account surface intensification of turbulent kinetic energy due to wave breaking, the turbulent length scale related to wave height is \( \ell = \alpha \xi \), the drag coefficient is \( \sqrt{C_D} = \frac{\alpha v (1-\beta_v)}{\xi^\beta_v} \), the eddy viscosity is \( v = \alpha v u_* \xi^{\beta_v} \), and the dissipation rate of kinetic energy enhanced by wave breaking is \( \varepsilon = \frac{\alpha e u^3}{H} \). \( \xi \) is a non-dimensional depth which is the depth divided by significant wave height (\( \xi = z/H \)) and \( \alpha_v, \alpha_e, \beta_v, \beta_e \), and \( \beta_\ell \) are empirical dimensionless parameters which are depth-dependent. To calculate depth-dependent eddy viscosity, the calculation of \( \beta_v = \frac{\beta_\ell + 4 \beta_v}{3} \), which is calculated from \( \beta_\ell \) and \( \beta_v \), is essential. \( \beta_\ell \) is 0 for \( \xi \ll 1 \) as the depth is less than significant wave heights and \( \beta_\ell \) is 1 for \( \xi \gg 1 \) as the depth is larger than significant wave height. \( \beta_v \) is \(-2 < \beta_v < -3\) (Drennan et al., 1992; Anis and Moum, 1992; Terray et al., 1996; Gerbi et al., 2009) in WASL (Wave affected surface layer), possibly dependent on wave age while \( \beta_\ell \) is -1 beneath WASL. The other parameters for calculation of eddy viscosity is \( \alpha_v = \alpha'_v W_0^\frac{1}{2} \) and \( \alpha'_v = S_M (B \alpha'_e \alpha'_\ell) \). The \( \alpha_\ell \) and \( \alpha'_e \) are an order-one fitting constant depending on the wave ages suggested in Terray et al. (1996). The empirical dimensionless parameter \( S_M \) is about 0.39, \( B \) is about 16.6 (Craig and Banner, 1994) and \( \alpha_\ell \) is an order-one fitting constant (~0.1). The wave age related constant \( \alpha'_e \) is assumed as 0.3 ~0.5 for young wave (Terray et al., 1996) but it was parameterized as 0.13 in the negative along-Sound wind while 0.2 in the positive along-Sound wind in western Long Island Sound (See Chapter 4). The \( S_M, B \) and \( \alpha_\ell \) vary with a slope value of eddy viscosity and \( \alpha'_e \) varies with wave age. The \( S_M (~0.39) \) \( B (~16.6) \) and \( \alpha_\ell (~0.1) \) are adopted from the previous
parameterization results taking into account turbulent wave breaking (Craig and Banner, 1994; Terray et al., 1996) but they are modified as $\alpha'_{e} \sim (0.1)$, $S_{M} \sim (0.2)$, $B \sim (7)$ and $\alpha_{e} \sim (0.4\sim0.9)$ for western LIS.

Figure 6. Surface eddy viscosity in the dependence of wind stress. Surface Stokes drift is taking out from the surface current.

Figure 6 shows the surface eddy viscosity estimated with observed surface shears in response to along-Sound wind using a simple gradient transfer method after taking out surface Stokes drift (Ch 3). The significant wind stress ($\sim 0.1$Pa), the averaged significant wave height, WBL thickness, WASL thickness and surface eddy viscosity from Figure 6 are used from the observed data and estimated values in the previous chapters for modeling and are summarized in Table 1. The Wave breaking Layer (WBL) is defined as the region where wave breaking transfers
<table>
<thead>
<tr>
<th></th>
<th>Negative $\tau_x$</th>
<th>Positive $\tau_x$</th>
</tr>
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<tbody>
<tr>
<td>$H_s$ (m)</td>
<td>0.3</td>
<td>0.2</td>
</tr>
<tr>
<td>WASL thickness (m)</td>
<td>5.8</td>
<td>4.6</td>
</tr>
<tr>
<td>WBL thickness (m)</td>
<td>0.75</td>
<td>0.45</td>
</tr>
<tr>
<td>Surface eddy viscosity(m$^2$/s)</td>
<td>0.0063</td>
<td>0.003</td>
</tr>
</tbody>
</table>

Table 1. The averaged significant wave height ($H_s$), WASL thickness, WBL thickness and the averaged surface eddy viscosity in WASL are shown to use for making the depth-dependent eddy viscosity profiles.

Figure 7. The vertical structure of eddy viscosity indicating BBL(Bottom Boundary Layer), SBL (Surface Boundary Layer), WASL(Wave Affected Surface Layer) and WBL(Wave Breaking Layer) and the black represents the constant eddy viscosity, blue represents log law eddy viscosity and red represents surface wave induced eddy viscosity models and
turbulent energy from the atmosphere to ocean. The Wave Affected Surface Layer (WASL) is where the near surface wave breaking is affected by turbulent kinetic energy rather than shear production. The Surface Boundary Layer (SBL) and Bottom Boundary Layer (BBL) are defined as the region where rigidity of the surface and bottom boundary produce the shear so that the distance of the boundary is proportional to the shear in the boundary layer. They are indicated in Figure 7.

Figure 7 also shows how the three different eddy viscosity profiles are different such that the black represents constant eddy viscosity profile, the blue represents log law eddy viscosity profile and
the red represents surface wave eddy viscosity profiles. The vertical structure of eddy viscosity in Wave affected surface theory is structured using Henderson’s parameterization (Henderson et al., 2013). Figure 8 describes the vertical structure of eddy viscosity profiles with three different eddy viscosity theories in the negative along-Sound and positive along-Sound wind using the information from Table 1. The dashed black line represents the constant eddy viscosity, the magenta line represents the log law (Rigid Boundary) eddy viscosity and the red line represents the surface wave eddy viscosity in the positive along-Sound wind. The black solid line indicates the constant eddy viscosity, the cyan line represents the log law (Rigid Boundary) eddy viscosity and the blue line represents the surface wave eddy viscosity. The averaged eddy viscosity of surface layer is about 0.0063 m²/s in the negative along-Sound and 0.003 m²/s in the positive along-Sound wind in surface wave eddy viscosity model as shown in Table 1. The depth-averaged eddy viscosity in the entire water column is about 0.0072 m²/s, i.e. 0.0065 m²/s in the positive along-Sound wind and 0.008 m²/s in the negative along-Sound wind. As surface eddy viscosity is asymmetric because of the asymmetric wave generation mechanism in negative along-Sound wind, the depth-averaged eddy viscosity is also asymmetric. This indicates that the surface mixing can influence not only in the surface layer but also in the entire water column.

5.5. Comparison of model and observation

Prescribing the three different depth-dependent eddy viscosities of Figure 8 in the model (Equation (11)), the wind and wave driven semi-analytical model is generated with finite element method and the model results are compared with the observation results in Figure 9. The blue line with the error bar in Figures 9 (a) indicates the wind driven currents in the negative along-Sound
wind while the red line with the error bar in Figure 9(b) represents those in the positive along-Sound wind from the observation. The grey dashed lines, the grey solid line and black solid line represent the model results with the constant eddy viscosity, Rigid boundary eddy viscosity and surface wave eddy viscosity respectively. As the observed results show the asymmetry, the model results also show the asymmetry with asymmetric eddy viscosity profiles.

Figure 9 (a) and (b). Comparison of observation and models. 9 (a) shows the along-Sound velocity in response to negative along-Sound wind and 9 (b) shows the along-Sound velocity in response to positive along-Sound wind. Red and blue lines represent observation results with error bars. The grey dashed line, grey solid line and black solid line represent the model results with constant eddy viscosity, rigid boundary eddy viscosity and surface wave eddy viscosity respectively.
In the wind-driven barotropic analytical model known as Winant’s model, axial, lateral and vertical circulation are shown in the positive axial wind with the rotation factors $\delta^{-1}$ which is the inverse non-dimensional Ekman number and represents the importance of rotation from Coriolis force in the various basin shapes (Winant, 2004). Winant model (2004) results prove that the rotation factor is important and can significantly affect the circulation along with various basin shapes. In the western Long Island Sound, the rotation factor $\delta^{-1}$, which is the ratio of the maximum depth to Ekman depth, is $\sim 2$ in the negative along-Sound wind while it is $\sim 2.6$ in the positive along-Sound wind in the significant wind event ($\sim 0.1Pa$). As Winant model (2004) shows the maximum value of rotation factor is about 6, the rotation factor is about 30% of maximum value in the negative along-Sound wind and about 40% of maximum value in the positive along-Sound wind in western LIS. This result show that as the mixing rate is more in negative along-Sound wind, the rotation factor decreases because the surface turbulence force competes with the rotation effect by Coriolis force.

The axial current in Winant model (2004), which is the same as the along-sound current here, is downwind in the surface layer and upwind below it. This is consistent with observation and model results with the three different eddy viscosity models showing that the downwind currents appear in the surface and upwind currents appear in the subsurface. The observation results show the upwind currents begins at 4~5m in the subsurface layer which is most consistent with the model results using surface wave eddy viscosity both in negative and positive along-Sound wind.

Figure 9 (a) and (b) also show how the vertical structure of eddy viscosity changes the circulation of the water by wind, wave and Coriolis forces. Although the depth-averaged eddy viscosity values in three different models are almost same, i.e. 0.0065 m²/s in the positive along-
Sound wind and 0.008 $m^2/s$ in the negative along-Sound wind, the vertical structure of currents are different from Winant model (2004) results, which are the same as constant eddy viscosity model results here, and the results with log law eddy viscosity model and Surface Wave eddy viscosity model. The results show that surface wave eddy viscosity model results are most consistent with the observed data while the rigid boundary eddy viscosity model taking into account the log law in the boundary is most inconsistent with the observation results in all cases because the water surface is not rigid and free and fluctuated by the complicated momentum and energy transfer.

5.6. Summary and Conclusions

Winds and waves along with tidal, buoyancy and Coriolis forces are the major forces driving water circulation in coastal basins. As previous chapters show, the surface currents, shear, waves, eddy viscosity and dissipation rate are asymmetrically generated by wind in a fetch-limited coastal basin such as western Long Island Sound. This asymmetry results from the different length of wind-fetch in the different wind events because of semienclosed and elongated basin shape. When the wind is from the east, the wind-fetch length increases along with the increase of the significant wave height which makes the surface shear decrease and turbulence mixing occurs more by wave breaking on the surface. This surface turbulence influences the circulation over the entire water column. Comparisons of the observed mean along-Sound patterns in response to different winds with the results of a wave and wind-driven semi-analytical barotropic model which is newly modified from Winant’s wind-driven barotropic model (Winant, 2004) to include wave dynamics and three cases of depth-dependent eddy viscosity are presented. These three different
eddy viscosity models show how the wind and wave-induced surface energy affect the dynamics in and below the surface layer.

The three different eddy viscosity models are constant eddy viscosity over the vertical, a rigid boundary eddy viscosity model and surface wave affected eddy viscosity model. The constant eddy viscosity model holds that the vertical mixing rate over the whole water column is constant. It is generally used for the classical Ekman dynamics to calculate Ekman number and Ekman depth (Ekman 1905) and also used in the wind-driven barotropic analytical model (Winant, 2004). The rigid boundary eddy viscosity model generates the eddy viscosity profile taking into account log law boundary layer near surface and bottom which is classical boundary layer theory (von Kármán, 1930). This model assumes that the boundary surface is rigid and that the distance from the boundary is proportional to the average velocity of a turbulent flow. For the interpretation of TKE balance in the rigid boundary case, the dissipation rate is balanced mostly by shear production. For the surface wave affected eddy viscosity model, the eddy viscosity profile is a function of surface wave induced kinetic energy transport due to wave breaking on the ocean surface. In other words, the surface dissipation of energy is mostly in balance with turbulent kinetic energy rather than the shear production in TKE budgets (Terray et al., 1996). So, the observed surface eddy viscosity in WASL (Wave Affected Surface Layer) is estimated using the simple gradient transfer method and then used to parameterize the structure of vertical eddy viscosity for various wind events. This surface eddy viscosity is estimated from the observed surface shears (Chapter 3) and the thickness of WASL is estimated from the observed wave height and wave age parametrization. (Chapter 4).

The depth-averaged eddy viscosity in the entire water column in western Long Island Sound is about 0.0072 m²/s, i.e. 0.0065 m²/s in the positive along-Sound wind and 0.008 m²/s in the negative along-Sound wind. These values can be compared to the eddy viscosity values from
the observation and model studies in Long Island Sound. Bennet et al. (2010) shows the eddy viscosity in western Long Island Sound is about 0.009 m²/s taking into account the tidal effect using the empirical equation by Bowden and Fairbairn (1952), i.e. $A_z = 2.5 \times 10^{-3} h_0 U_t$ where $h_0$ is the maximum depth and $U_t$ is the tidal velocity. Whitney and Codiga (2011) takes 0.003 m²/s for western Long Island Sound and 0.012~0.03 m²/s for eastern Long Island Sound to simulate the wind-driven circulation for Long Island Sound. They comment that it is difficult to set the appropriate eddy viscosity for the simulation. About the depth-averaged eddy viscosity in western Long Island Sound, the value I use in this chapter is smaller than Bennet et al. (2010) which takes account tidal effect and is larger than Whitney and Codiga (2011). Recently, Schwendeman et al. (2015) shows the integrated surface dissipation rate is about $4 \times 10^{-4}$ m³/s³ when the wave breaking begins to occur. Transferring the eddy viscosity from the dissipation rate with their results using Equation (9) in Chapter 4 ($\varepsilon = \frac{15}{2} \nu \cdot \frac{r}{\rho_0}$), the averaged eddy viscosity is about 0.0077 m²/s which is almost consistent with the average surface eddy viscosity (0.0072 m²/s) described in Table 1.

Comparing observation and model results, the vertical structure of currents in response to each wind event show the downwind currents near surface and upwind currents in the subsurface. The observed surface data is from high frequency radar while those in the deeper water column are from the bottom-mounted ADCPs. The averaged rotation factor in western LIS is about 2.3 which is 38 % of maximum rotation values which indicates the rotation factor is moderately important (Winant, 2004). This is consistent with the range (1.7~3.8) that Whitney and Codiga (2011) calculated for Long Island Sound.

In conclusion, first, surface wave turbulent energy significantly affects circulation in coastal basins and should be included in the momentum equation. To explain and to predict surface water dynamics, wave effects along with wind force is found the most important. Second, the
vertical structure of eddy viscosity taking into account ocean surface wave breaking is useful to understand ocean surface boundary dynamics and to predict more realistic circulation patterns in the near surface. The depth dependent eddy viscosity is newly suggested with three different profiles and used to prescribe in the new momentum equation. This permits explanation of how the water mixes and moves not only in the surface layers but also at depth in response to winds and waves. As the observations and the results of the wind and wave driven barotropic semi-Analytical model applied to western Long Island Sound have shown, it seems clear that surface turbulence through wave breaking is interactive near surface in the coastal ocean.
Chapter 6. Summary and Conclusions

Although surface wave dynamics are important in determining coastal ocean circulation along with wind, tide, Coriolis and buoyancy forces, the observation analyses and modeling studies have ignored the effect of the surface wave in the momentum balance. Especially, since the geography of western Long Island Sound results in a fetch-limited basin where wind wave generation and circulation is influenced by the asymmetric wave field response to differing wind directions, surface wave dynamics must be taken into account to understand how the water mixes or moves.

The statistics of wind fluctuations in Long Island Sound are summarized. Observations of significant wave height and the relationship to wind are used to show how the wind and surface waves are linked in a complex, fetch-limited coastal basin. Wind wave parameterization that depends on fetch is assessed. I show that the wave field in the western Long Island Sound has an asymmetric response to wind direction. The fetch-dependent models are consistent with the observations except for times when the wind is from the east. I conclude that more complex physics taking into account the local geography, i.e. bottom dissipation effect, depth, width of the sound, must be considered to explain the observations.

The critical results are shown to explain the important role of the wind and wave-generated turbulence on the surface currents. The asymmetric surface currents, surface shear and the estimated eddy viscosity depending on wind direction and magnitude are presented using observation data from CODARs (Coastal Ocean Dynamics Application Radar) and ADCPs (Acoustic Doppler Current Profiler). These are caused by asymmetric wave dynamics such that the wind-fetch lengths are different asymmetrically in the various wind events, especially in the
complicated fetch-limited coastal basin; western Long Island Sound. Surface eddy viscosity is estimated using simple gradient transfer method with surface shear and surface Stokes drift which is included in CODAR velocity estimates is calculated. After taking out Stokes drift from the CODAR surface data, the asymmetric mixing rate induced by wave are demonstrated with larger eddy viscosity (up to 5 times) when the wind is negative along-Sound than positive along-Sound. I concluded that the asymmetry of the surface eddy viscosity and surface Stokes drift results from the asymmetric wave dynamics and the surface Stokes drift is an important factor modifying the surface shear asymmetry.

Furthermore, the thickness of Wave Affected Surface Layer (WASL) and Wave breaking layer (WBL) are estimated in order to describe the depth to which the wave breaking affects the water column. The asymmetric surface eddy viscosity and dissipation rate are estimated with wave-induced method (Henderson’s parametrization, Henderson et al. 2013) taking into account surface wave breaking in order to show how much the surface turbulent energy is dissipated in response to various winds. This wind and wave-induced asymmetry in the surface layer indicates that near-surface intensified turbulent energy is transferred from the wind field to the water column through wave breaking at the surface. The wave steepness is shown to determine the critical wind stress to occur the wave breaking and the result shows wave breaking significantly occurs when the wind stress is larger than 0.1 Pa in western Long Island Sound. Using the Henderson’s parameterization, $\alpha$, which is explain the ratio of wave speed from wind and from the ocean, is parameterized as 0.1 in the negative along-Sound wind and 0.2 in the positive along-Sound wind for western Long Island Sound.

Lastly, these observational results are compared with model results to show how wind and surface waves affect circulation in the coastal ocean. Comparisons between the observed wind-
driven currents and results of a wave and wind-driven barotropic semi-analytical model modified from Winant’s analytical model (Winant, 2004) with the addition of a depth-dependent eddy viscosity term are presented. I use three different eddy viscosity models with the three different theoretical approaches, a constant eddy viscosity model, a Rigid boundary eddy viscosity model and a surface wave-induced eddy viscosity model. In the results, the model with surface wave-induced eddy viscosity is most consistent with the observations. So, I conclude that the surface turbulent wave dynamics significantly influence the coastal ocean circulation and depth-dependent eddy viscosity term should be included in the momentum equation at least for western Long Island Sound.
Appendix A: Analytical Solution (Winant, 2004)

Winant (2004) shows non-dimensional continuity and momentum equation and the equations are

\[ \nabla \cdot u + w_z = 0 \quad (a) \]

and

\[ \frac{2}{\delta^2} k \times u = -\nabla \cdot \eta + \frac{\partial^2 u}{\partial z^2} \quad (b) \]

Defining that the complex variables as velocity \( U = u + iv \) and the pressure gradient \( N = \eta_x + i\eta_y \), Equation (b) is rewritten as

\[ \frac{d^2 U}{dz^2} - \frac{2i}{\delta^2} U = N \quad (c) \]

with the boundary conditions are \( U = \tau_x \) at \( z=0 \) and \( U=0 \) at \( z=-h \), the analytical solution of Equation (c) is

\[ U = \tau_s \frac{\sinh[\alpha(z + h)]}{\alpha \cosh(\alpha h)} - \frac{N}{\alpha^2} \left[ 1 - \frac{\cosh(\alpha z)}{\cosh(\alpha h)} \right] \quad (d) \]

Where the surface stress \( \tau_s = \tau_x + i\tau_y \) and \( \alpha^2 = 2i/\delta^2 \).
Appendix B: Finite element method (Galerkin’s method)

The non-dimensional momentum equation from Equation (11) of Chapter 5 with depth-dependent eddy viscosity is rewritten in the function of $z$ as

$$K_z U''(z) + K'_z U'(z) - \frac{2}{\delta^2} K_z U(z) - K_N = 0 \quad (a)$$

Defining the complex basis functions as

$$\phi_0(z) = T(z - 1) \quad (b)$$

$$\phi_1(z) = z(z - 1) \quad (c)$$

where the value of $U$ at $z=0$ is $T$.

The first function $\phi_0(z)$ is a straight line between the boundary points assuming a linear relation between boundary conditions

$$\phi_0(z) = U_a + (U_b - U_a) \frac{z - a}{b - a} \quad (d)$$

where $a$ is $z$ at $z=0$ and $b$ is $z$ at $z=1$, i.e. $a=0$ and $b=1$, and $U_a$ is $U$ value at $z=0$ $U_b$ is $U$ value at $z=1$. Note that $U_a$ is $T$.

$T$ is determined using Euler’s method and a shooting technique (Press W. H. et al. (1992)).
Using the given the point of \( U_a = T \) and \( U_b = 0 \) when \( z = 0 \) and \( z = 1 \) and \( a = 0 \) and \( b = 1 \), Equation (d) becomes Equation (b).

Then, we can make the trial solution

\[
U(z) = \phi_0(z) + C \phi_1(z)
\]  

(e)

The trial solution \( U(z) \) can be rewritten as

\[
U(z) = C z^2 - (T + C)z + T
\]  

(f)

where complex value \( C \) is the coefficient.

\( U'(z) \) and \( U''(z) \) are then

\[
U'(z) = \frac{1}{2} C z - (T + C)
\]  

(g)

\[
U''(z) = \frac{1}{2} C
\]  

(h)

Putting Equations (f), (g) and (h) into Equation (a), the residual function \( R(z) \) is

\[
R(z) = K_1 \left( \frac{1}{2} C \right) + K_2 \left( \frac{1}{2} C z - T - C \right) + K_3 \left( C z^2 - (T + C)z + T \right) + K_4
\]  

(i)
where the complex values $K_1 = K_z(z)$, $K_2 = K'(z)$, $K_3 = -\frac{2i}{\delta z}K_z(z)$ and $K_4 = -NK_z(z)$.

This residual function is used to construction $U(z)$ so that the integral will be zero for the weighted function.

The integral of the inner product of the residual function with trial function is

$$\int_0^1 \phi_1(z) \cdot R(z) \, dz = 0 \quad (j)$$

The integral of Equation (j) in terms of $C$ is

$$\int_0^1 \phi_1(z) \cdot R(z) \, dz = C^2 \left( \frac{1}{5}K_3 + \frac{1}{8}K_2 - \frac{1}{2}K_3 + \frac{1}{6}K_1 \
- \frac{1}{3}K_2 - \frac{1}{6}K_2 - \frac{1}{4}K_1 - \frac{1}{2}K_2 + \frac{1}{2}K_3 \right) + C \left( -\frac{1}{4}K_2 T - \frac{1}{3}K_2 T + \frac{1}{3}K_3 T \
+ \frac{1}{3}K_4 - \frac{1}{2}K_2 T + \frac{1}{2}K_3 T - \frac{1}{2}K_4 \right) - 0 = 0 \quad (k)$$

Simplified Equation (k) to get the coefficient $C$ with coefficients $A$ and $B$ is

$$AC^2 + BC = C(AC + B) = 0 \quad (m)$$

where $A = \frac{1}{5}K_3 + \frac{1}{8}K_2 - \frac{1}{2}K_3 + \frac{1}{6}K_1 - \frac{1}{3}K_2 - \frac{1}{6}K_2 - \frac{1}{4}K_1 - \frac{1}{2}K_2 + \frac{1}{2}K_3$ and $B = -\frac{1}{4}K_3 T - \frac{1}{3}K_2 T + \frac{1}{4}K_2 T + \frac{1}{2}K_3 T - \frac{1}{2}K_3 T - \frac{1}{2}K_4$ from Equation (k).

Using the given $A$ and $B$, the coefficient $C$ is calculated as
Finally, Galerkin’s solution $U(z)$ is obtained with the calculated coefficient $C$ in Equation (f).

\[ C = -\frac{B}{A} \]
Bibliography


